THE UNIVERSITY OF ALBERTA

THE ICE IN COULTHARD CAVE

BY

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The undersigned certify that they have read, and recommend to the Faculty of Graduate Studies for acceptance a thesis entitled The Ice in Coulthard Cave submitted in partial fulfilment of the requirements for the degree of Master of Science.

Abstract

Some features of Coulthard Cave situated near the Crowsnest Pass, Alberta, Canada, are not explained by the basic theories of the formation of ice in caves. The complete extent of this cave is impossible to discover because the known limits of the passageways are completely blocked with relatively clear ice.

Temperatures below 0°C obtained in late summer indicate that the cave probably remains below freezing throughout the year. The formation of cold air traps due to the geometry of several passages likely aids, but is not essential for, the persistance of ice. Accumulation of ice at depth in the passages formed blockages. Solidification of ponded water is inferred from the vertical orientation of large crystal optic axes. Later, a supercooling of the water at depth produced protocrystalline aggregates containing numerous air and sediment inclusions.

At present, the rate of ice ablation is approximately 3mm annually, and this is inferred from an indicator deposited on the ice. Ablation produces surface scallops bounded by sediment capped interstitial ridges. Adhesion of the sediment to the ice, and movement by normal trajectory, account for the transportation of ice sediment inclusions to concentrations. The sediment composition of the bed rock, ice inclusions, and surface material is relatively uniform.

Oxygen isotope analysis indicates that the ice probably did

not form when large global scale ice masses existed during glacial times, and likely formed before glacial times or since the Hypsithermal Period ended some 4,000 years before the present.

Passages of most other known ice caves in the Rocky Mountains contain only crystal formations and ice stalagmites. Because Coulthard Cave is atypical, further research is suggested.

ACKNOWLEDGEMENTS

My deepest gratitude is expressed to Dr. M.C. Brown without whose supervision and kind assistance this work would not have been possible.

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P. Smart and G. Pilkington of the University of Alberta served as willing porters for equipment and samples. Mr. R. de Saussure of Cave Research Associates, California, presented numerous suggestions while visiting Coulthard and in later correspondence. Many individuals connected with the McMaster Karst Research Group supported the work in Coulthard. In particular R. Bignell and T. Morris spent long hours in the chill of the cave, the while offering inspiring commentaries. The original cave survey was also completed by the McMaster Group. Thanks is extended to all for coming to Coulthard, often not for your pleasure but for my safety.

My sincerest thanks to Ethel for her invaluable criticism of the text.

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Contents

Figures

Introduction

The general purpose of this work is to present some techniques for the investigation of ice caves[1], and to describe their application to a unique cave situation. Articles have been published in America (Halliday, 1954; Merriam, 1950) and more extensively in Europe (Abel, 1953; Gèze, 1965; Trombe, 1952) but these are mainly descriptive in nature. The lack of study may be accounted for by the wide and complex variety of circumstances under which ice forms within caves.

All but the main passage of Coulthard Cave, Mount Coulthard, Crowsnest Pass, Alberta, Canada, end in ice blockages. Unlike many ice caves in which the ice is seasonal or takes the form of ice speleothems[2] or flows, that in Coulthard Cave completely blocks passages some 6 metres in diameter. The passage may be seen extending over 3 m beyond the surface of the extremely clear ice.

Exploration of the cave was undertaken by the McMaster University Karst Research Group in the summer of 1968[3]. Several climbs

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[1] Ice caves - "permanent caves in rock formations in which ice occurs and may remain into the summer or throughout the year." Henderson, J., "Caverns, Sink Holes and Natural Bridges." Colorado University Studies, 1932, Vol. 20, p. 115-127.

[2] "Speleothem" - A secondary mineral deposit formed in caves, such as stalactite or stalagmite. Synonymous with "formation".

[3] It should be noted that access to the cave, even in summer, is difficult and requires advanced rock and snow climbing technique.

leading to the ice blocked passages required extensive removal of loose material, indicating that they had not been previousely attempted. At this time, a penny was left on one of the ice surfaces as a possible indicator of ice accumulation or ablation. A temperature reading of - 3.0°C was obtained at one of the ice blockages.

Ice in caves is the product of long term climatic and micro-meteorologic factors. A cave can be considered as an unique natural laboratory (Poulson and White, 1969) in which a relatively constant environment permits long-term biologic and geologic experiments. It was hoped that crystallographic and isotope studies would reveal information about the age of the ice, under an initial hypothesis that the ice might be of glacial age. The relevance of these findings will be discussed below.

Chapter I

The Problem and Its Geographical Setting

Introduction

The Rocky Mountains of Canada, about 1,400 km long and 160 km wide, extend from the Liard River south to the United States border. Throughout, the rocks lie in a series of thrust sheets and folds, exposing one of the world's foremost limestone mountain systems. The great thicknesses of pure limestone[1], high degree of initial jointing, large amplitude of local relief (which generates steep hydraulic gradients), cold water enriched with CO_2, and erosion extending into the Pliocene, satisfy the requirements of the development of karst[2] landforms on a major scale.

Geomorphological studies undertaken in the region have been concerned mainly with glacial retreat; lately a small group has been interested in karst landforms. Even less work has been done on the association of karst and glaciology, even though the interrelationship was noted in 1907 (Canada, Department of the Interior). More recent discoveries, especially that in Mount Coulthard, indicate the existence

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[1] Limestone - 75 per cent or more $CaCO_3$ by weight (Goudge, 1944).

[2] Karst - A landform in which chemical solution by flowing water is the dominant geomorphic process.

of natural laboratories where certain aspects of the physics of ice may be studied over long periods in a partially controlled setting. The cave within Mount Coulthard is located at 49° 33' 53" N., 114° 34' 05"W., or some 9.6 km south southwest of Coleman, in the Crowsnest Pass region of Alberta, Canada (Figure 1).

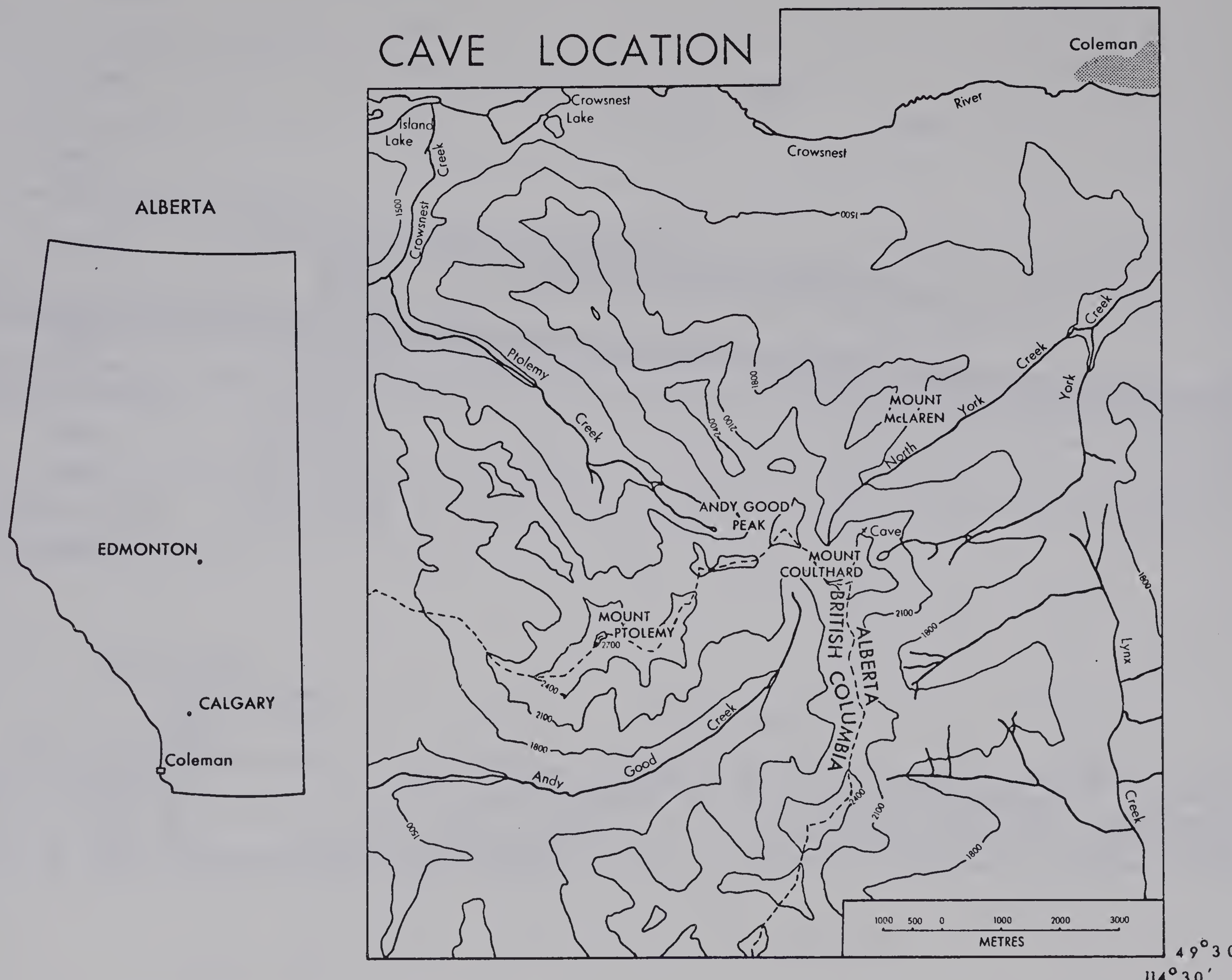
Geology

The various formations exposed within the Crowsnest area reveal the following points:

> (1) the alternation of marine and terrestrial deposits and the occurrence of volcanic deposits in the Upper Cretaceous; (2) a remarkable thinning of all the formations from west to east; (3) the absence of Tertiary formations; (4) the existence of three unconformities of considerable magnitude, (a) the disconformity indicated by Devonian sediments resting upon Middle Cambrian sediments..., (b) the erosional unconformity between the Palaeozoic and the overlying Mesozoic rocks, and (c) the erosional unconformity of less magnitude between the Kootenay and Blairmore Formations (MacKay, 1932, p. 32B).

The area between the Kootenay Lower Cretaceous sediment present in the immediate Coleman region, and the Rocky Mountains, is underlain by Alberta Group marine shales and Belly River freshwater sediments (Figure 2). These have a fairly uniform westerly dip, averaging 32°. The former are estimated to be approximately 914 m thick while the latter are some 1,220 m in thickness (Price, 1966, p. 67).

At Crowsnest Lake, to the west of Coleman, the Belly River Formation is over-ridden by the westerly dipping Palaeozoic limestones


ALBERTA
EDMONTON
CALGARY
Coleman
CAVE LOCATION
Coleman
Crowsnest
Lake
Island
Lake
Creek
Crowsnest
River
Crowsnest
1500
1500
Ptolemy
Creek
1800
2100
2400
MOUNT
McLAREN
North
York
Creek
York
Creek
ANDY GOOD
PEAK
Cave
MOUNT
COULTHARD
MOUNT
PTOLEMY
2700
2400
2100
1800
Andy
Good
Creek
BRITISH COLUMBIA
ALBERTA
2100
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2400
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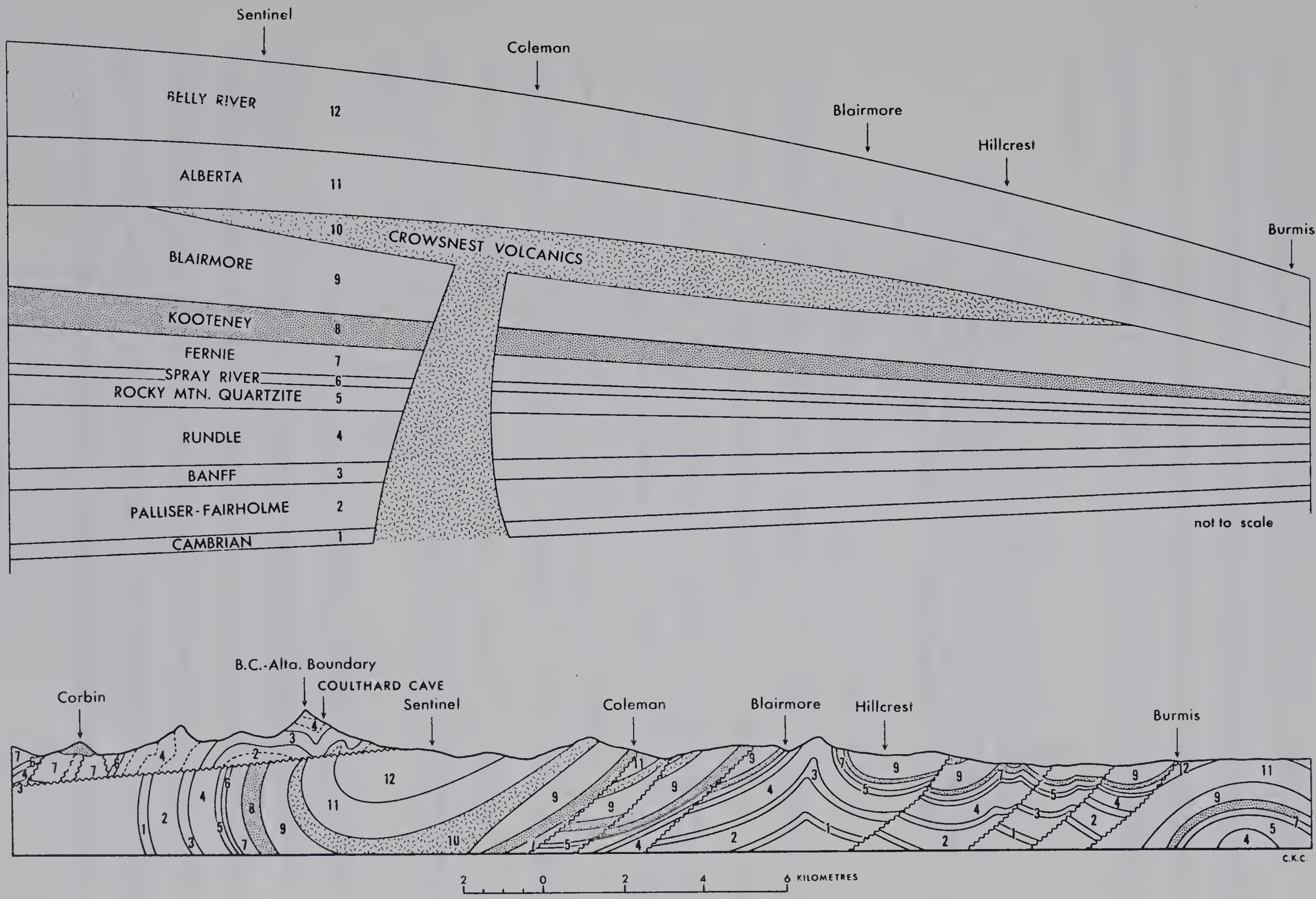
1500
1000 500 0 1000 2000 3000
METRES
49°30′
114°30′

Figure 1



Section from Corbin to Burmis before and after the Rocky Mountain revolution. (Mackay,1932,Figure 6)

Figure 2

that constitute the main range of the Rocky Mountains. The strata with a combined thickness of 3,000 m have a westerly dip of 45 to 60°. A few minor faults have caused slight displacement. The outcrop in the Crowsnest Pass has a surface width of 4 km. However, as MacKay has stated:

> ...five miles (8 km) to the south in the latitude of Mount Coulthard and Mount Ptolemy, where the Alberta-British Columbia boundary line trends in an east-west direction, the Palaeozoic rocks are folded and faulted, and the outcrop has a width of approximately 5 miles (8 km) (1932, p. 35B).

The Mount Coulthard area is comprised of three main stratigraphic units: the Palliser Formation, the Banff Formation, and of the Rundle Group, the Livingstone Formation. The Palliser Formation consists of dark grey, fine-crystalline limestone, commonly mottled and locally interbedded with brownish grey, medium-crystalline dolomite. These Devonian sediments are overlain by the Banff Shales of the Mississippian period. The strata are composed of dark brown and black limestones containing nodules and lenses of chert. The presence of chert and hydrogen sulphide suggests stagnant water conditions during formation (MacKay, 1931, p. 10B).

In turn the Banff shale is overlain, without any apparent break, by a thick series of massive, coarse and finely crystalline limestones of the Livingstone Formation in which the cave is located. On weathering, the colour of the rock shifts from the predominantly grey to white and buff. Nodules, lenses and irregular bands of white or grey chert up to three inches in thickness mottle the formation.

The surficial geomorphology of the Coulthard-Andy Good region is that of classical alpine glaciation. As many of the radiocarbon dates obtained in Alberta refer to the Keewatin glaciation, a series of dates from the Purcell Trench region of British Columbia may serve as a more accurate guide to the dating of the Wisconsin glaciation of the Crowsnest Pass. Dates GSC-740 and GSC-715 indicate that the area was not glaciated between 43,800 C^{14} years and 25,840 C^{14} years B.P. (Fulton, 1968, p. 1078). This interval coincides with the Sangamon Interstade - locally called the Olympia Interglaciation. Further dates reveal that the ice did not advance over the area until after 20,000 years B.P. and had retreated prior to 10,000 years B.P. (Fulton, 1968, p. 1079). In the Banff area of Alberta, final glacier recession predated 9,300 years ago (Denton, 1970, p. 34). Recorded fluctuations of alpine glacier systems during late Wisconsin recession appear not to have been contemporaneous on a regional basis (Fulton, 1968, p. 1080).

It appears that due to the development or expansion of cirques during the Wisconsin and previous glaciations, caverns such as Cleft Cave (Coward, Drake and Thompson, 1969, p. 49) and several caverns in Mount Coulthard have been truncated. These glaciokarst[3] features are also found in other parts of the Rocky Mountains.

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[3] Glaciokarst - A glaciated limestone region possessing both glacial and karst characteristics.

Holocene climate

While some discussion persists as to the rate of climatic amelioration subsequent to glacier retreat, several periods have been recognised in the literature. The early Hypsithermal Period, a slightly cooler and drier interval from 11,000 to approximately 10,000 years B.P. (W-2285), preceded a warm climate from 8,000 to 4,000 years B.P. New evidences from Europe and America suggest the warm period stimulated glacier growth, high timberlines and rapid soil formation in areas of summer rainfall, in the approximate range of 7,000 to 6,000 years B.P. (Curry, 1970, p. 25).

Other evidence for a warmer period has been obtained from soil studies and radiocarbon evidence of former forest advances approximately 5,500 and 900 years B.P. in Keewatin (Bryson, 1965, p. 47).

The oxygen^{18} isotope content of atmospheric precipitation may be utilized to study past climate, since the ratio of $O^{18/16}$ depends on the mean annual temperature of the site of precipitation (Dansgaard, 1953; 1954). Thus a 1°C change in mean annual temperature generally causes a much larger change in the $O^{18/16}$ ratio in atmospheric precipitation compared with the effect of a similar change of water temperature. The oxygen isotope record suggests that mean annual temperatures during the Hypsithermal Period were about 2°C higher than at present; temperatures some 10,000 years ago are estimated to have been 1-2°C lower than at present (Stuiver, 1970, p. 128).

Within the past several centuries, minor climatic fluctuations have been recorded by glacier movement.

> ...the Athabaska Glacier reached its maximum about 1714 and withdrawal began in different places in 1721 and 1744. It readvanced in the first half of the 19th century, reaching almost to its maximum extent. Withdrawal began between 1841 and 1866. Recession has continued with minor fluctuations, and recessional moraines were formed in about 1900, 1908, 1925, and 1935. Total recession from 1721 to 1953 was about 3,600 feet (1,100 m) (Mercer, 1958, p. 2b.2.10).

Meek has correlated the 1866 recession and several later fluctuations of Canadian glaciers with retreats in the Swiss Alps indicating widespread rather than localized variations (1948, p. 266).

Present climate

A sufficient distribution of stations with reliable records is essential for the construction of accurate temperature and precipitation profiles. In a mountainous area, stations located vertically as well as spatially are useful, but rarely available. Thus drawing a profile without such 'anchor' points is difficult and obten subject to large errors. Within the Crowsnest Pass only one station, Coleman, is noted in the published records of precipitation. Lundbreck, some 26 km to the east, measures both temperature and precipitation (Canada, Department of Transport, 1967).

The precipitation normal for Coleman is 505.7 mm annually, That for Lundb eck is 489.8 mm (Canada, Department of Transport, 1967). Both stations have minor peaks in the summer, though of the two, the

former is less pronounced. The mean annual daily temperature at Lundbreck is 2.9°C. Only seven months have averages above the freezing point. It is not surprising, therefore, to find that August temperatures in Coulthard Cave, located on a north facing slope, are below freezing, and thus permit the retention of ice throughout the year (Figure 1).

Speleogenesis[4]

The entrance to Coulthard Cave is located in limestone of the Livingstone Formation some 650 m above North York Creek; it may be postulated that speleogenesis initiated at a higher erosional surface. The elevation of 2,650 m above mean sea level is similar to the 2,320 to 2,620 m altitudes of cave entrances at the head of Ptolemy Creek and Andy Good Plateau (Coward, Drake and Thompson, 1969, p. 47). Numerous karst windows[5] throughout the region indicate previous cave systems now nearly completely eroded by glacial and fluvial action. The absolute height of the former erosional surface during the speleogenesis of Coulthard Cave may not be accurately determined.

Subsequent to initiation, solution by a down-cutting stream formed vadose trenches[6] (Gardner, 1935; Swinnerton, 1932; 1942), enlarging the cave to almost its present dimensions. The formation

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[4] Speleogenesis - The initial formation of the cave system.

[5] Karst windows - A small natural bridge or arch which can be seen through.

[6] Vadose - Passages formed by stream corrasion.

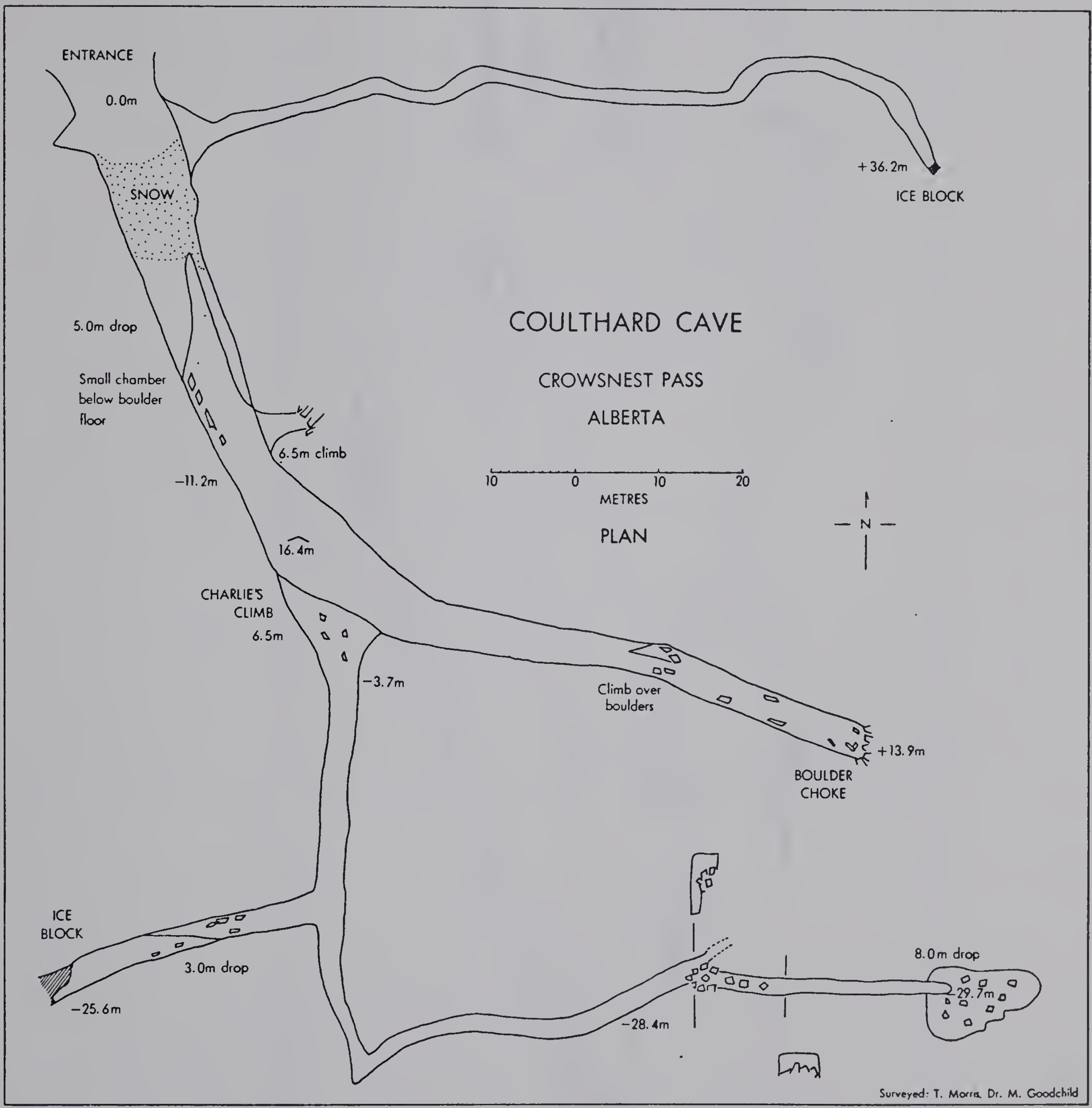
ENTRANCE
0.0m
SNOW
5.0m drop
Small chamber below boulder floor
6.5m climb
−11.2m
16.4m
CHARLIE'S CLIMB 6.5m
−3.7m
COULTHARD CAVE
CROWSNEST PASS
ALBERTA
10 0 10 20
METRES
PLAN
N
+36.2m
ICE BLOCK
Climb over boulders
+13.9m
BOULDER CHOKE
ICE BLOCK
3.0m drop
−25.6m
−28.4m
8.0m drop
−29.7m
Surveyed: T. Morris, Dr. M. Goodchild

Figure 3

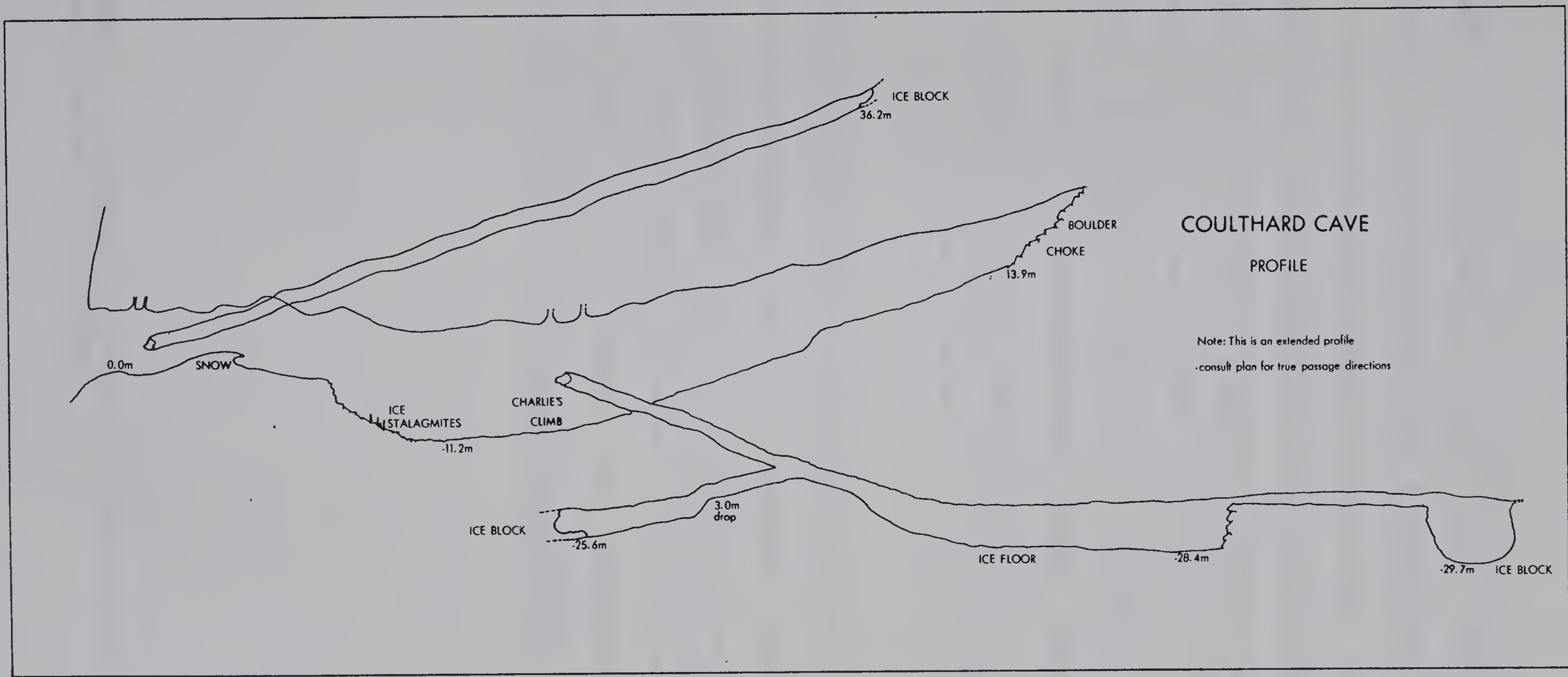
COULTHARD CAVE
PROFILE
Note: This is an extended profile
-consult plan for true passage directions
ICE BLOCK
36.2m
BOULDER
CHOKE
13.9m
0.0m
SNOW
ICE
STALAGMITES
-11.2m
CHARLIE'S
CLIMB
3.0m
drop
ICE BLOCK
-25.6m
ICE FLOOR
-28.4m
-29.7m
ICE BLOCK

Figure 4

of speleothems under conditions similar to those in some of the caves of Vancouver Island and northern California, with a temperature range of 7.2 to 10.0°C, occurred throughout much of Coulthard Cave. These calcite deposits may date from warm periods of either the Pliocene or the interglacials of the Pleistocene. Extensive deposition has also been recorded in Ice Hall, Ptolemy Creek region (Coward, Drake and Thompson, 1969, p. 45), and to a lesser extent in Gargantua Cave, Andy Good Plateau. Climatic conditions which were probably warmer resulted in some re-solution, but little passage down-cutting, and this concluded the period of active speleogenesis (R. de Saussure, Pers. comm.).

Frost shattering contributes to the present floor of massive breakdown[7]: no evidence of glacial polish is illustrated. Because of the shattered rock, passages which branch off from the high walls of the main cave require special climbing techniques. For this reason, and because of undisturbed dust and sediments on the passage floors, it is certain that the cave was not fully explored until 1968.

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[7] Breakdown - Enlargement of parts of a cave system by fall of rock masses from walls and ceiling.

Chapter II

Characteristics of Ice Caves

Definition

Natural rock caves containing ice, commonly termed ice caves have been defined by Henderson as, "...permanent caves in rock formations in which ice occurs and may remain into the summer or throughout the year (1932)." The term, therefore, excludes caves dissolved into ice, such as may occur in glaciers ('glacier caves'). The study of ice formed within caves may correctly be included within, and use the techniques natural to, glaciology, because "glaciology is the study of frozen water in any of its forms or locations (Pounder, 1965, p. 9)."

In contrast to the high-temperature minerals, ice seldom finds thermodynamic conditions corresponding to its stable state. Only in the frozen zone of the lithosphere can it be preserved for a geologically long period of time. Even at negative temperatures, ice in the atmosphere experiences constant change under conditions of atmospheric saturation by water vapor, while on the earth's surface, the texture of ice is subject to thermal influences. Regions of constant temperatures below 0°C are rare on the surface of the earth, with the present intensity of solar radiation (Shumskii, 1964, p. 11). Even in arctic regions, surface ice may be subject to, and altered by, temperatures above freezing.

Thus caves have attracted the attention of investigators for the study of crystals formed by the sublimation of water directly from gas to solid state. Under conditions of slow growth, Stulov (1949) studied these crystals in Russian caves, while Al'tberg, Troshin and Golovkov (1939) described the Kungur ice cave in the Urals. Sublimation crystals may vary in size and shape over periods of several years, and, since no trace of ablated material is left, these crystals are not accurate, long term indicators of previous climates.

Ice caves

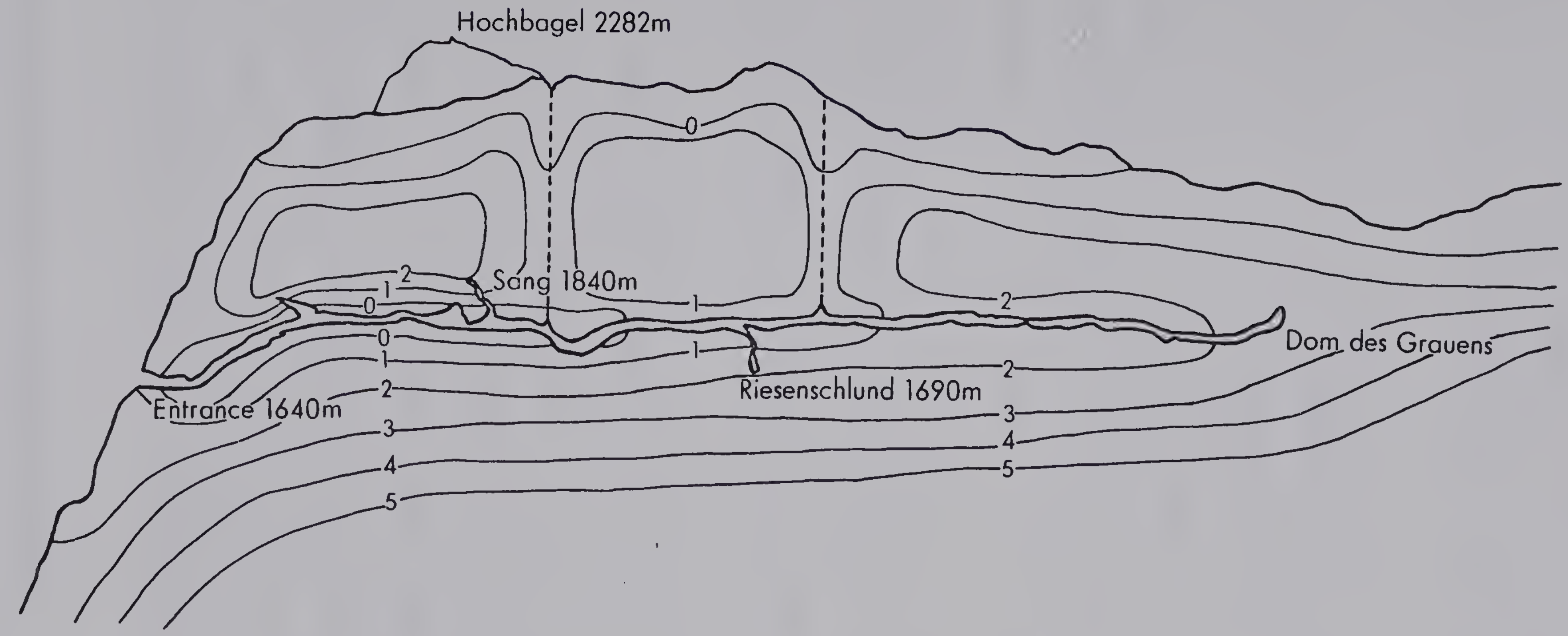
Many known ice caves, including Coulthard Cave, require some source of moisture and temperatures below 0°C for the formation and persistence of ice. Deposition can be direct from gas to solid states if the air is saturated with respect to ice in the immediate vicinity of the point of deposition, or by the simple freezing of liquid water. Thus the fundamental variables are temperature and moisture conditions, and both of these are strongly dependent on cavern climate and air circulation. While climate determines whether or not and how much ice is formed, a feedback mechanism exists because the deposition of ice may have a significant influence on the cave climate.

Climate changes are damped out by the confining high heat capacitance of rock, though quite marked variations occur in many caves both in time and location. The shorter period fluctuations in external atmospheric conditions are damped out rapidly, so that while diurnal

fluctuations in climate external to a cave may not be transmitted far into the interior, seasonal variations frequently are (Cropley, 1965). These latter changes determine whether or not ice is deposited (Stupishin, 1959).

The air circulation to and from, and within, a cave is one of the most important factors controlling the cave climate especially if that flow is turbulent (Wigley and Brown, 1971). Air can move readily through caves with multiple entrances at different levels by simple gravitational drainage (Abel, 1953; Geze, 1965, p. 128), or by external atmospheric pressure fluctuation. Thus the temperature of cave walls is determined by changes in air temperature rather than, as is more commonly supposed, vice versa (Brown, Ford, and Wigley, 1970).

Ideally several classes of caves may be distinguished by the type and location of ice deposits within them: entrance deposits, regions with low air flow, and points along passages with high air flow (Brown, Ford, and Wigley, 1970). It is possible, however, for individual caves to exhibit more than one of the characteristic depositional forms (Trombe, 1952). An example of the first general group is Eisriesenwelt, in Austria, which contains extensive permanent ice deposits only in the first one kilometre of the 40 km cave system (Geze, 1965, p. 117) (Figure 5). Humidification of air is endothermic: the air from outside entering the cave slowly increases its humidity, consequently cooling. If the incoming air becomes saturated, then excess water vapor turns into liquid phase and condenses, or, under negative temperature conditions,



Geoisotherm diagram of Eisriesenwelt Cave, Austria (Oedl, 1923).

Figure 5

it may sublimate directly into ice (Stelcl, 1965, p. 175). The sublimation of atmospheric water vapor to solid state, or the freezing of liquid water near the entrance will liberate heat and tend to raise the air temperature deeper inside the cave (Oedl, 1923; Brown, Ford, and Wigley, 1970).

An implication of entrance ice deposition in a model cave is that there will tend to be periods of accumulation and ablation during any one year. Over a number of years, though, the ice deposits will tend to be in equilibrium, otherwise growth would eventually block or nearly block the entrance part of the cave and this would cause a considerable effect on the cave climate in a direction towards reducing the accumulation and the establishment of an equilibrium state. Such a regulatory effect damps the magnitude of climate changes which might otherwise occur.

The feature illustrated in Figure 6 is an extensive snow and ice deposit in the entrance of Coulthard Cave. In cold summers, saturation of the snow by surface melting proceeds slowly, and mainly ice crystals with less than 5 mm diameters are formed. Some firn[1] may be left at the end of the summer. In warm summers, saturation of the snow cover is complete and water may stand in a pool at the foot of the ice to form very coarse ice when it refreezes (Figure 6). Attempts to determine

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[1] Firn - Recrystallized and compacted snow.

Figure 6. The entrance snow deposit as seen looking into the cave, upper photo, and the melt water pool on the entrance side of the snow deposit (R. de Saussure photos).

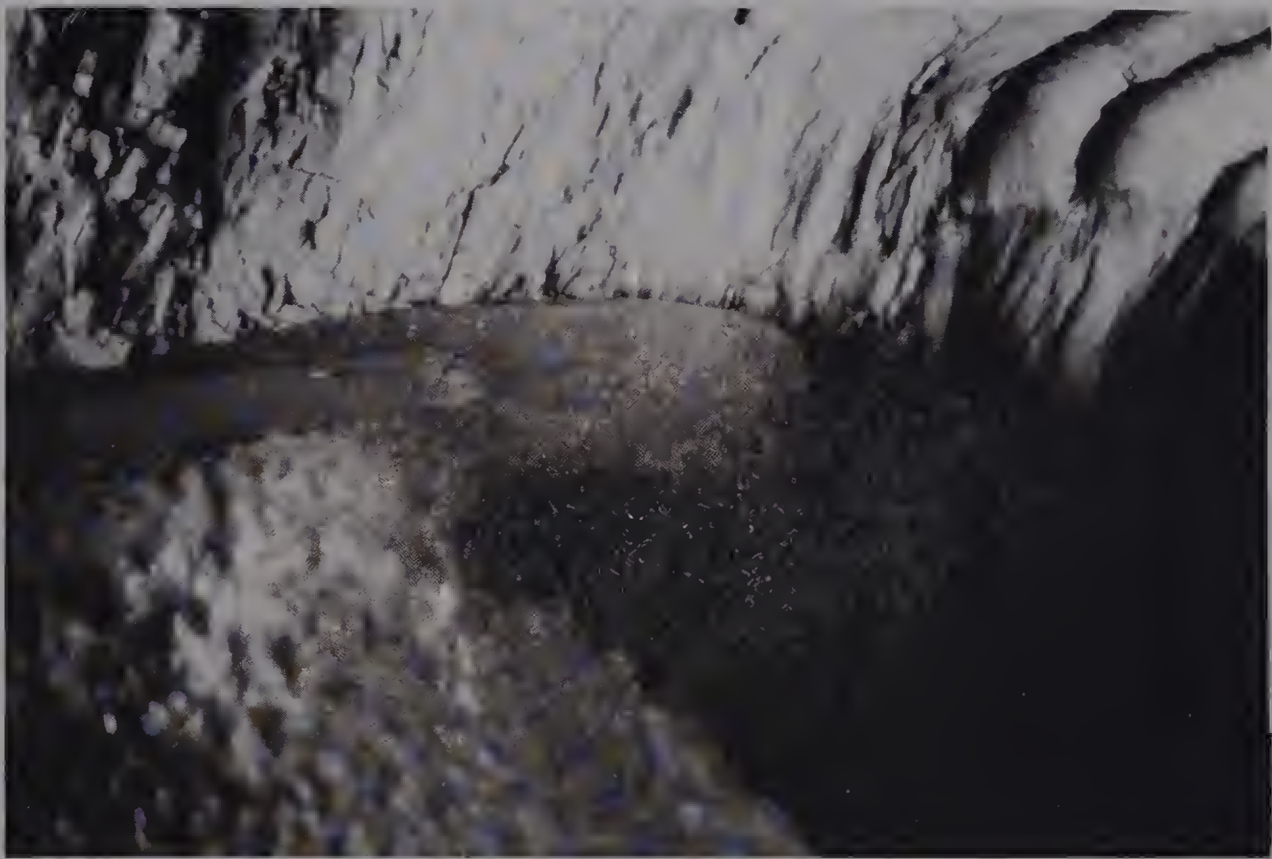

Figure 7. The snow deposit cornice indicating winter wind direction into the cave, now undergoing summer ablation (R. de Saussure photo).

variations of cave climate from the entrance snow and ice deposit must incorporate the following facts: (1) summer ablation may eliminate one or more depositional horizons, and (2) variations of the deposits are likely affected by climatic cycles with periods of the order of 10 years. Therefore the snow near the cave entrance was not subject to detailed analysis.

The inner portion of the entrance snow deposit takes the form of a cornice (Figure 7), indicating that the dominant winter storm wind direction is into the cave, exiting at the end of the main passage through a boulder choke[2]. A reverse flow occurs in summer, when cool air exits the cave via the main entrance. This is simply seasonal reversal of air circulation (Geze, 1965, p. 128-130). Further into and below the snow deposit are numerous ice stalagmites some 75 mm in diameter and over 0.6 m in height formed by the accumulation of seepage moisture falling from the roof and freezing on the floor or on previous ice deposits.

When air circulation in a cave of small size is very low, ice deposition occurs in the furthest reaches of the cave (Pirker, 1929; Trombe, 1952). The dominant ice form is a sublimation deposit; clusters of hexagonal ice plates develop on the roof and walls when the temperature is -0.5 to -5.0°C (Stupishin, 1959). For large crystal growth, a very stable environment is required, but some air circulation

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[2] Boulder choke - A passage completely blocked by rock debris.

is essential for moisture to be transported into these parts of the cave. In Coulthard Cave, crystals reach a diameter of 12 mm on the roof of 'Charlie's Climb'. More extensive deposition occurs at aven[3] entrances directly above the entrance to the climb (Figure).

A third type of ice cave occurs when high velocity air circulations are present in a cave. Rapid, complex, or even rapidly changing air circulation may enable ice to form in regions remote from the entrance zone along main passages rather than side chambers containing still air.

Ice at depth in Coulthard Cave is found primarily as blockages at the ends of passages, and occasionally on the floor of the cave. The geometry of the passages often slopes downward, and this will "trap" pools of cold air (Geze, 1965, p. 129). The ice is of high clarity with inclusions in the form of vertical bubble trains. "Scallops"[4] are the dominant surface morphology: interstitial ridges are capped with sediment.

Theories of Coulthard Cave ice origin

Several theories including both glacial and non-glacial processes have been suggested to explain the ice in Coulthard Cave. Little

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[3] Aven - A vertical or highly inclined shaft in limestone extending upward from a cave passage.

[4] Scallop - Round or oval hollow in the ice surface often having an asymmetric cross section along its main axis.

scientific work has been undertaken on cave ice masses because mainly these formations are seasonal rather than perennial in nature. The massive size of the Coulthard Cave ice, however, prompted suggestions of glacial origin from either plastic squeezing into the cave or else a water filled cave which froze. Plastic flow of ice into the cave would likely have removed projections of rock, smoothed the passage walls and left distinctive grooving. It would be expected that ice crystals would parallel the long axis of the cave. None of these glacial features are found within the cave.

Solidification of a completely water filled cave would exert a considerable pressure on the passage walls because of the coefficient of expansion of ice. Any cracks and joints in such a cave would be expanded by the freezing water. Now, during ablation of ice in the main passage, extrusion of pressurized ice in cracks of the rock walls should result. These also were not observed in Coulthard Cave.

Present ice may have developed in the post glacial period from the ponding of seepage water by blockages in the cave passages. Under this hypothesis ice crystals are expected to be similar to those formed in surface ponds, with long vertical crystals growing down from the surface. It is possible that the ponded water could have been formed during the Altithermal Period by the melting of Wisconsin age ice deposits. While the characteristics of the ice may be similar, determination of the osygen isotope ratios should indicate more positive values for the recent cave seepage than for melt water.

At present snow is found only in the first 32 m of the cave

even though dominant winter wind direction is into the cave. It is doubtful, therefore, that the relatively weak draught in the side passages could transport sufficient snow to block the passage. Also, sufficient pressure for recrystallization of the snow would be absent.

Both petrographic and oxygen isotope studies were undertaken to determine likely modes of ice formation within the cave.

Chapter III

Crystallographic Research

Introduction

Crystallographic research provides the basis from which petrographic study of ice may be undertaken because, "...properly speaking, modes of occurrence are determined by geological, not petrographic methods...(Shumskii, 1964, p. 107)."

The aims of crystallographic studies of ice are to establish the geometric relationships of crystals of hexagonal symmetry forming ice, and of the orientation of the crystallographic or c-axes with respect to the crystal form, bed position or foliation, and direction of movement, if any. Ice is unstable (Holden, 1963) and difficult to preserve; thus, as a rule, crystallographic studies are carried out in the field or in field laboratories. The need for equipping the laboratory with a good source of light creates a number of difficulties, only partially alleviated by the relative ease of sample collection and the constant negative temperature of the cave. Logistical constraints and the availability of permanent laboratory facilities elsewhere prompted field sampling: analysis was undertaken wholly in the laboratory.

There are three principal methods of studying the normally transparent crystals of ice. The first is selective melting of the ice to determine the crystal boundaries. This technique involves the melting down of an ice surface, stopping periodically to measure the crystal

diameters. This method was not used because crystal diameters could be measured on thin sections of ice cut for determination of crystal c-axes. The second method, based on optical crystallography, permits not only the study of the crystal configuration and the orientation of the c-axes, but also reveals the presence of internal crystal stresses. The third method, x-ray analysis, permits accurate study of the spatial orientation and the space lattice of the crystals, but is unable to show the crystal geometry. Furthermore, x-ray studies require complex equipment and special training.

Ice petrography

Petrography is the detailed study of the texture and the structure of rocks. The composition of ice is simple, comprising, at most, four phases: ice crystal, and inclusions of water, air, and minerals. Thus, the description of an ice aggregate consists mainly of the delineation of the texture: the geometry, relative amount, form and orientation of the component parts. The following elements enter into the concept of the texture of a monomineralic rock such as ice:

(1) the external form and size of the crystals;

(2) the spatial relationships of the crystals;

(3) the relationship of the crystallographic orientation to the external form of the crystals;

(4) the relationship of crystals to the inclusions;

(5) the relation of the crystallographic orientation to the bedding elements or the regularity of the texture (Shumskii, 1964, p. 104).

In petrography, the structure of rocks is distinguished from texture in that structure defines the spatial disposition of component parts and the extent to which the rock fills space. In particular, the structure treats the stratification, jointing, schistosity, cleavage, and porosity of the rock (Bader, 1951, p. 520).

The study of the internal structure of ice is made easier by its optical properties. Clear ice, that is ice free of inclusions[1], is transparent to visible light. Optically, ice is a uniaxial crystal, the optic axis being the c-axes of crystallographic notation. Transmission of light is normal when the rays are incident parallel to the c-axes. If the light is incident at an angle to the c-axes then it is separated into rays which travel at different speeds and so are refracted at different angles (Pounder, 1965, p. 16). Using two polaroid filters with their planes of polarization at right angles, almost no light is transmitted through the ice, permitting determination of the c-anes of single crystals.

Ice petrography may reveal not only the crystal geometry, presented immediately below, but also the orientation pattern of an ice aggregate which reflects the pattern of stress and strain in ice.

<u>Ice samples</u>

Because of the inconvenience of working at low temperatures

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[1] Inclusions - Foreign matter contained within ice. Inclusions may be any or all of solid, liquid, or gaseous states.

in an isolated location, oriented samples were removed from the ice blocked southwest passage at -25.6 m in Coulthard Cave (Figures 8 and 9). Even though most temperate glacier ice contains numerous bubbles of trapped air (Flint, 1947, p. 15), the literature contains references that ice "...as free from air bubbles as possible...(Ahlmann, 1949, p. 269)" should be sought for petrographic analysis, and "...sufficiently weathered to render the depressed crystal boundaries visible to the eye ...(Seligman, 1949, p. 256)." It is only recently that ice containing gas inclusions has been studied (Coachman, 1956; Jonas and Muller, 1969; Matsuo, 1966; Scholander et al, 1956; 1958; 1960).

Two blocks of ice each approximately 300 mm on a side were orientated then removed from the cave. Block A was from the main ice deposit in the passage which contains numerous air inclusions. Block B was obtained from a recess in the wall. Here the ice is almost free of inclusions, the few present are concentrated in trains. Figure 9 shows the location of the blocks in the passage.

Crystal size

To give a quantitative definition of the external form and size, and the spatial relationships of the crystals, volumetric measurement was undertaken. Several methods of determining size, with varying degrees of accuracy may be utilized. While the simplest method consists in measuring the largest and smallest diameters of the plane cross section of the crystals, error exists in that a section will not pass through the largest diameters of all the crystals. For small

Figure 8. The removal of ice samples in the southwest passage of Coulthard Cave.

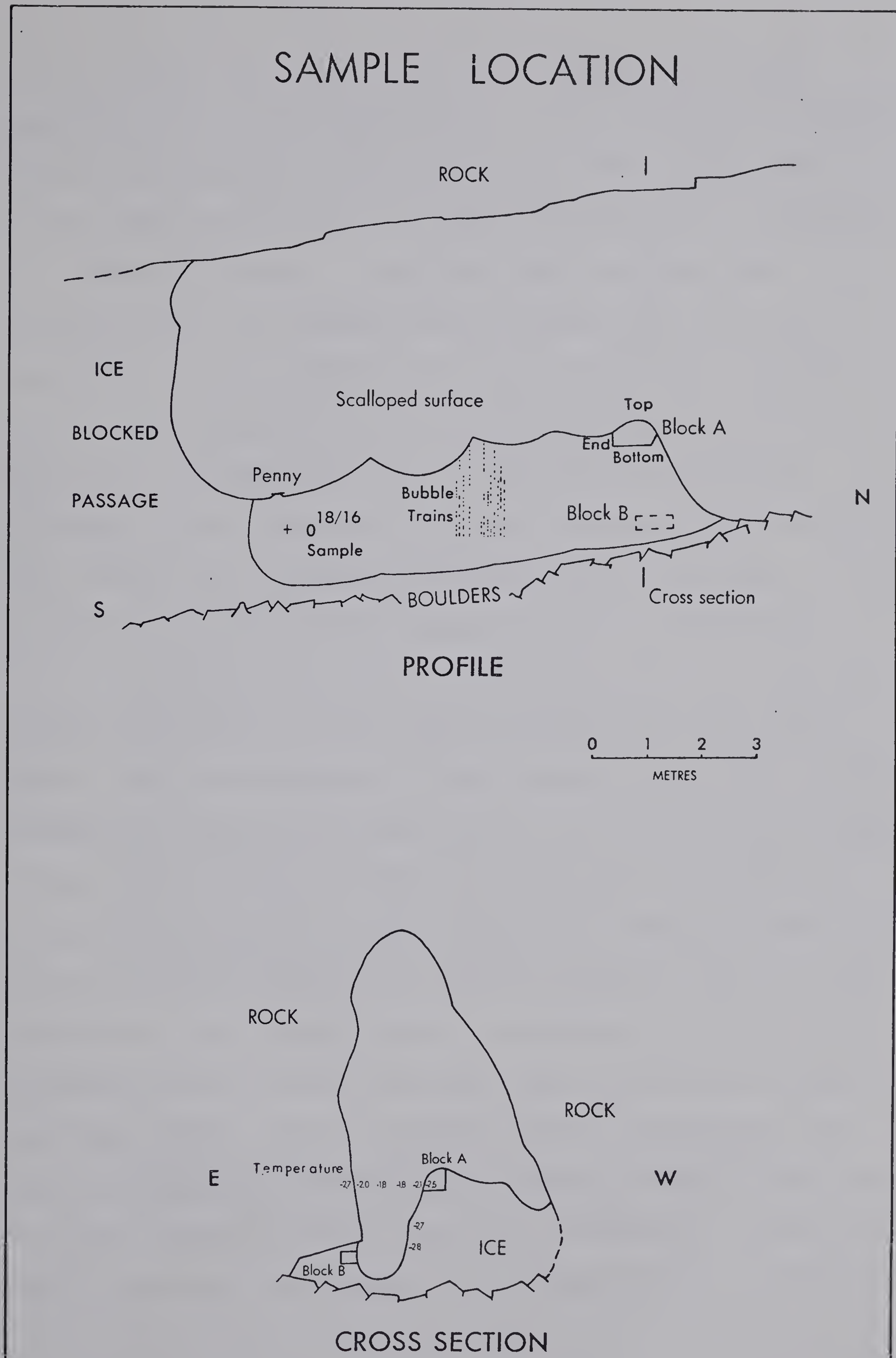
SAMPLE LOCATION
ROCK
ICE
BLOCKED
PASSAGE
Scalloped surface
Top
Block A
End
Bottom
Penny
Bubble
Trains
Block B
N
+ 0 18/16
Sample
S
BOULDERS
Cross section
PROFILE
0 1 2 3
METRES
ROCK
ROCK
E
Temperature
-2.7 -2.0 -1.8 -1.8 -2.1 -2.5
Block A
W
-2.7
-2.8
ICE
Block B
CROSS SECTION

Figure 9

diameter crystals this is not a very serious error as they measure roughly the same on their three axes: a series of circles with logarithmic progression in diameters may be used (Ahlmann, 1949; Johnson, 1946; Seligman, 1949). Each crystal is classified according to the standard size nearest it in area. A greater source of error exists when measuring crystals of rectangular section (Figure 10). Planimetric determination of each crystal is a more accurate method in that case.

The accumulation curve of distribution of the total volume of the crystals of different sizes is plotted on the basis of the percentages of area, with processing of the data done by the Johnson (1946) method, i.e., according to the percentages of the volume occupied by the crystals of each group in the sample (Figure 12).

Determination of the mean size of the grains was affected by several concentrations of large numbers of grains less than 1 mm^3 in volume. This notwithstanding, the mean volumes are 12.6 and 5.1 mm^3 for the upper and lower portions respectively of block A. The long hexagonal columnar crystals comprising some 5 per cent of the total number of crystals are confined to the upper portion of the block. They are greater than 75mm in length and the ratio of length to width is in the order of 8:1. Their original length is impossible to determine because of ablation, however, several taper with depth, indicating previously much larger volumes. Others are parallel columns: interspersed in the crystal boundaries are isolated or, more commonly, trains of small and very small crystals (2-5 mm and less than 1 mm diameter). Crystals of 1 mm^3 or less tend to be spheroids, while examples in the 1 to 16 mm^3

South

North

Figure 10. The end section of block A, illustrating the marked difference in crystal size. The crystal volumes were determined from within the enclosed area.

Figure 11. The basal thin section of block A. The enclosed area is the lowest plane of measurement of crystal volumes. North is parallel to the vertical edges of the enclosed area.

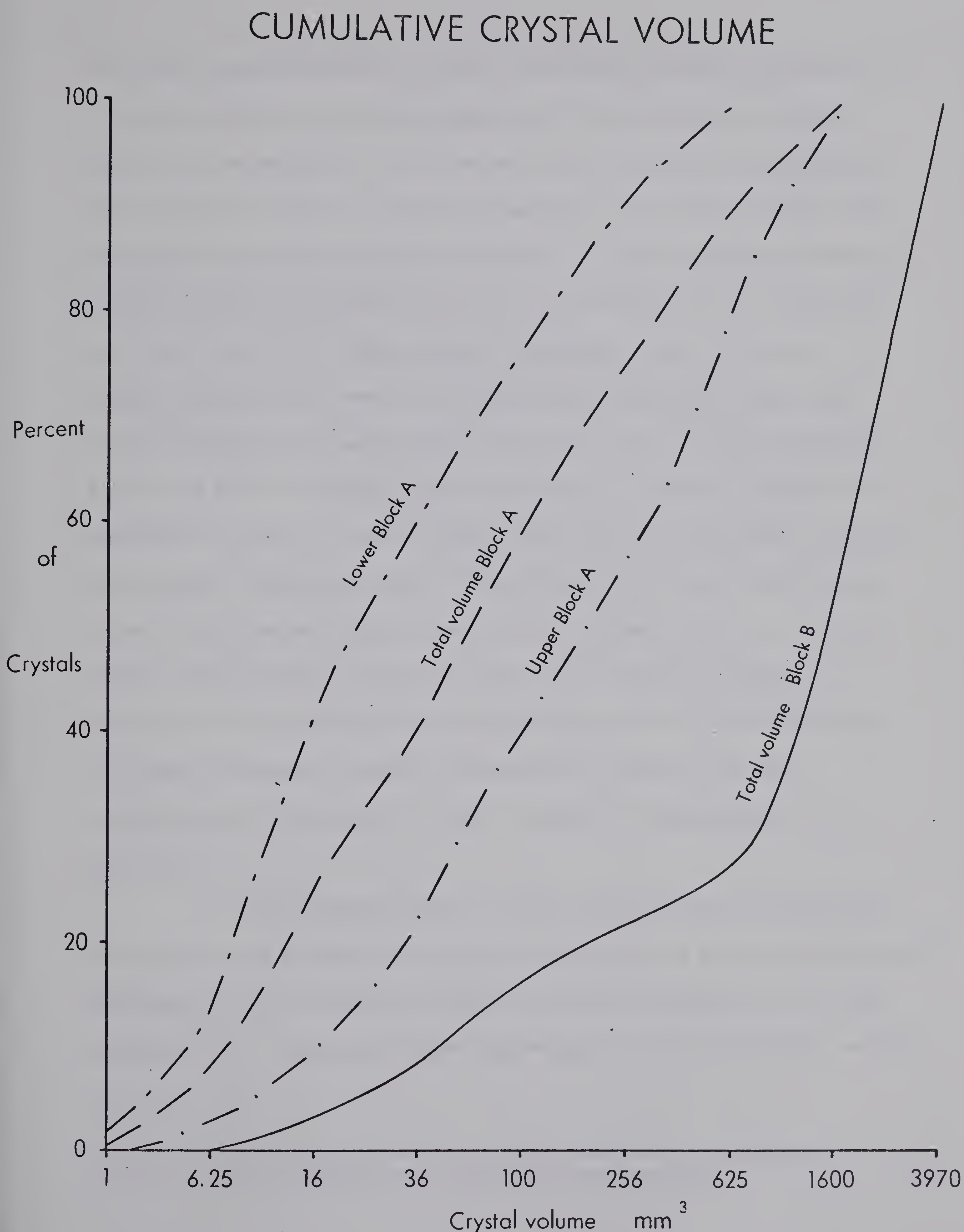
CUMULATIVE CRYSTAL VOLUME
100
80
60
40
20
0
Percent
of
Crystals
1
6.25
16
36
100
256
625
1600
3970
Crystal volume mm3
Lower Block A
Total volume Block A
Upper Block A
Total volume Block B

Figure 12

range have protruberances at random orientation giving an appearance of a highly crenated, interlocking aggregate. The alternating layers of grains with diameters of 2 to 5 mm and 6 to 16 mm and, in particular, the elongation typical of foliated glacial ice are absent (Figure 13). The whole of the upper section of block A - large crystals, columns of smaller crystals, and gas inclusions - is orientated with a strike of 152° and a dip of 10°. The passage orientation is 64°. Below the evident dividing line some 90 mm below the crest of the block, the texture changes to crystals whose three axes vary in length less than a ratio of 2:1. No apparent shear distortion or diminant elongation of crystals is evident. A band of less than 1 mm to 5 mm diameter crystals does persist immediately below the dividing line, extending down some 40 mm. The occasional large crystal with a diameter over 16 mm may be noted. Significantly, the band of less than 1 to 5 mm diameter crystals is also the area with the greatest number of gas inclusions, the number decreasing somewhat downwards and almost completely disappearing in the region of large crystals in the upper half of the block.

It should be mentioned that the very presence of impurities, which impede the growth of ice crystals, facilitates protocrystallization[2]. Experiments in the growing of large ice crystals showed that the new crystals have a tendency to appear opposite the points of bubble contact

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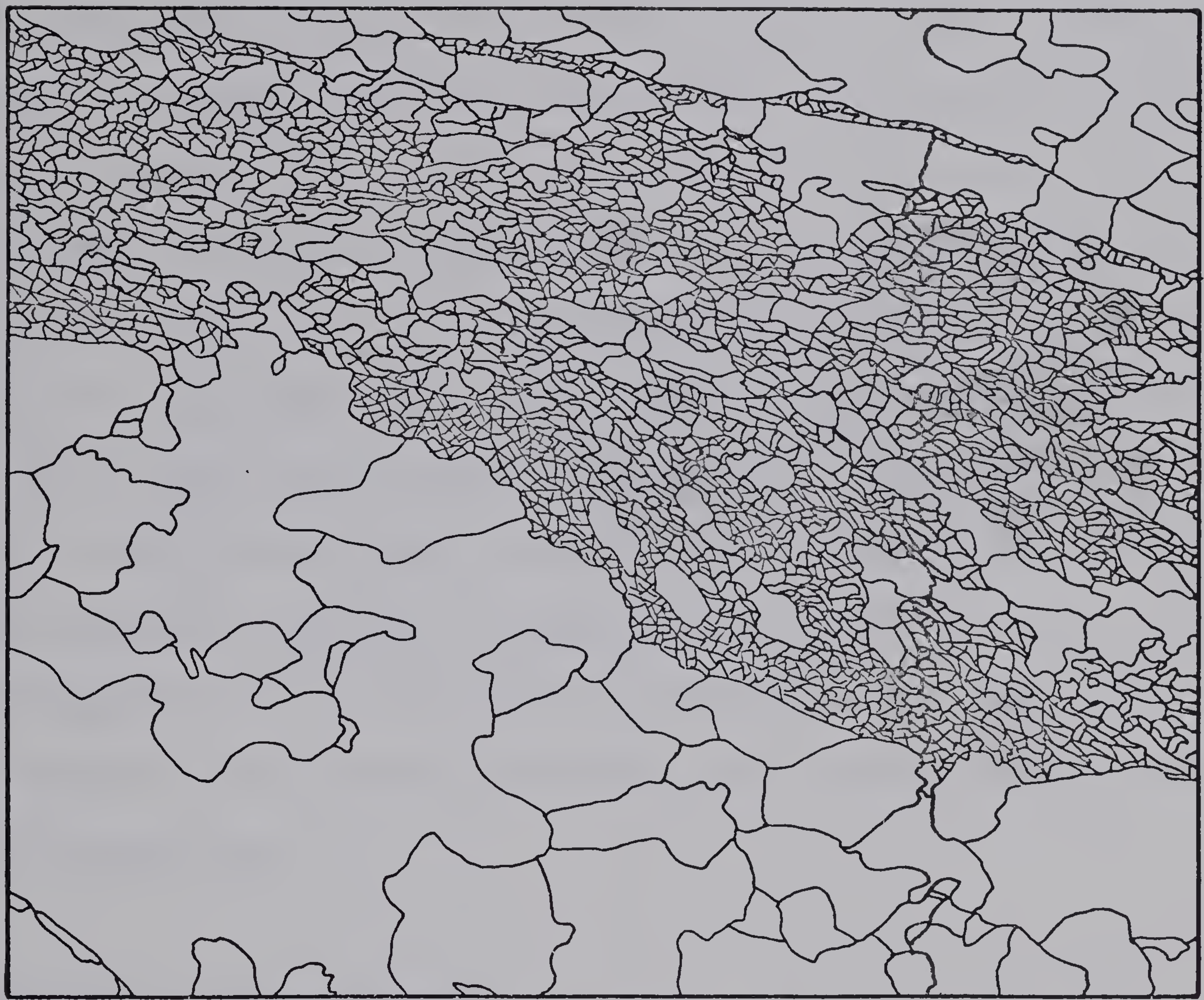
[2] Protocrystallization - first, original, or simplest crystallization initiated on a substance other than ice.



Crystal distortion in glacier flow (x 1)

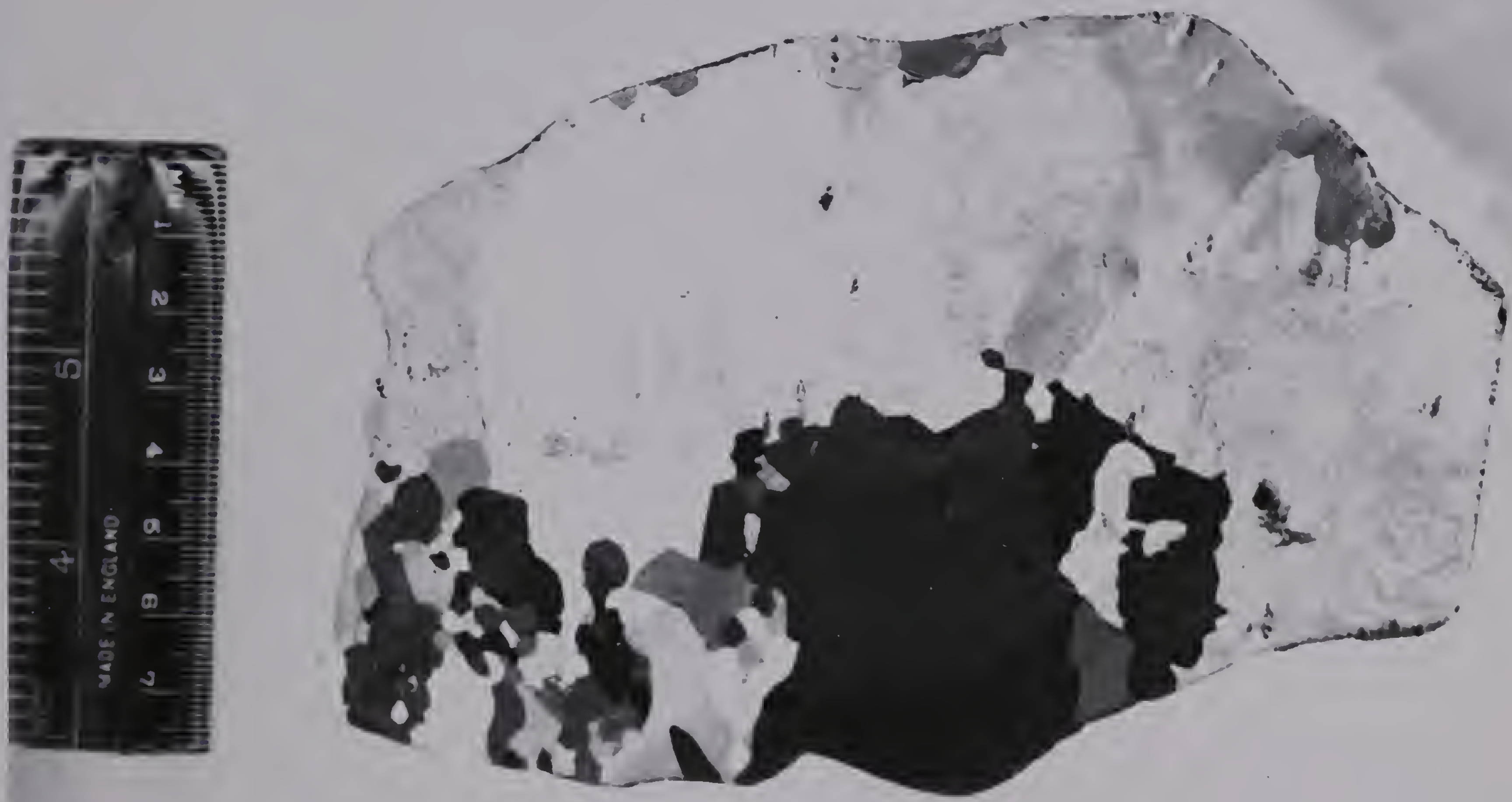
Figure 13

with the original crystal (Adams and Lewis, 1934).

Determination of the mean crystal size for the whole of block B was possible because of the few grains forming that block. The inordinate size of one crystal which comprises some 63 per cent of the basal area, skews the graph of cumulative volume. In fact, the maximum diameter is somewhat larger than those found in many glacial dead ice deposits (Ahlmann, 1949, p. 274)(Figure 14). Some 18 per cent of the total number of grains are greater than 16 mm in diameter while those of less than 1 mm are, on the other hand, lacking. The 6 to 16 mm size grains, here associated with vertical columns of gas enclosures, are very well rounded: sharp corners and protrubances common to block A are seldom visible. Interestingly, crystal stress is easily seen in the larger crystals of block B. The phenomenon is exhibited in only two of the elongated crystals in the upper portion of block A. There is no correlation of stress and the relatively small, isolated bubble trains in the latter block.

Orientation of the optic axes

Knowledge of how individual crystals form and their interaction with surrounding crystals having different optic orientations can be developed from crystal petrofabric studies. The fabric of a crystalline mass is found by the statistical study of the orientation of the individual crystals and their relationship to each other (Rigsby, 1960, p. 589). For example, Rigsby observed a direct relationship between crystal orientation and the foliation planes in his studies of the Saskatchewan



West East

Figure 14. The inordiante size of several crystals of block B are apparent in horizontal sections.

Glacier, Alberta (1954), and of the Moltke Glacier, Thule, Greenland (1955). He reported the strength of the pattern to be more or less proportional to the strength of the shearing stresses imposed on the ice. It is only by the study of the optic axes that the orientation may be discerned, for in spheroidal crystals the geometry is seldom controlled by the orientation and, hence, by the basal plates of growth.

The samples used for fabric studies were the same as those utilized in determining crystal sizes. It was necessary, however, to ascertain and record the geographical orientation of the blocks prior to their removal from the cave so that no errors in the thin section orientation during laboratory study would occur.

The laboratory cold room[3] is similar to the design outlined by Bader (1951). In the room an electric bandsaw was used to trim the blocks of ice and to cut thin sections to a thickness of 2 mm. Mounting of the thin sections onto glass slides was done on a smooth surface, steel hot plate. Photographic equipment was kindly made available so that permanent records of the ice thin sections under polarized light could be obtained. The ice crystal c-axes were measured on a universal stage fabricated by the Department of Civil Engineering.

In the laboratory, the selected directions on which to cut thin sections included the horizontal plane for the base, a long axis parallel to the passage orientation for the side, and a third at right

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[3] The laboratory cold room was kindly made available by Dr. J. Nuttall, Department of Civil Engineering, University of Alberta.

angles to the other two forming the end. From the data thus obtained, it is possible to draw up projections in the three dimensions.

A universal stage utilizing essentially the same principles as those first introduced by Fedorov in 1891, and standard ice petrofabric techniques (Bader, 1951; Langway, 1958; Lewis, 1960) were used for the crystallographic studies. Lewis obtained an accuracy of $\pm$ 2° (1960, p. 27), while Langway believes that the total error in analysis is no more than 5° (1958, p. 8).

Fabric diagrams - stereograms - were made by plotting the c-axes on the lower hemisphere of a Schmidt equal-area projection in the conventional manner (Bader, 1951; Haff, 1938; Langway, 1958; Lewis, 1960; Phillips, 1960; Shumskii, 1965) and were orientated so that they represented horizontal sections with north at the top. Although some thin sections were made from vertically cut slabs because of greater ease and accuracy in cutting, the diagrams have all been rotated so that they may be projected on a lower hemisphere (Phillips, 1960). In studies of metamorphic rock fabrics, the optic axes of 400-500 grains are frequently plotted on one diagram. In most cases, ice fabrics are strong so that one hundred grains will easily reveal the pattern at any one sample location (Rigsby, 1960). Langway (1958) suggests that at least two hundred axes should be plotted in order to obtain a reliable statistical analysis, although, he adds that if the section indicates a very strong pattern, then it is possible to use fewer points. Petrofabric analysis carried out for this study on block A entailed determination of 234 axes. Only some one hundred crystals in

block B were determined because of the limited number of crystals comprising the block.

Interpretation of the data from the fabric diagram of block A indicates density concentrations as high as 7 per cent in 1 per cent of the net area. This maxima is adjacent to the primitive[4] at an orientation of 110° while the passage bearing is 64° (Figure 15). A nearby significant trend crossing the stereogram on an axis of 150 — 330° is associated with the large crystals in the upper half of the block. The trend consists of some 41 per cent of the total number of crystals. The isolated 5 per cent area of the net at the centre of the diagram is orientated along this axis at right angles to the 64° bearing of the cave passage. A second maxima containing some 27 per cent of the crystal c-axes is located parallel to the passage direction. As noted above, no foliation or similar planes were observed in the ice.

Detailed crystallographic analysis of block A reveals, therefore, that the density concentrations are much weaker than glacial fabrics mentioned by Rigsby (1960), and that the stronger of the two maxima is normal to the bearing of the cave passage.

The fabric diagram derived from block B (Figure 17) is similar to the one of Block A, in that it also exhibits a centre concentration as well as a strong maxima normal to passage orientation. In this case, the trend is approximately 140 - 320°. as part of the

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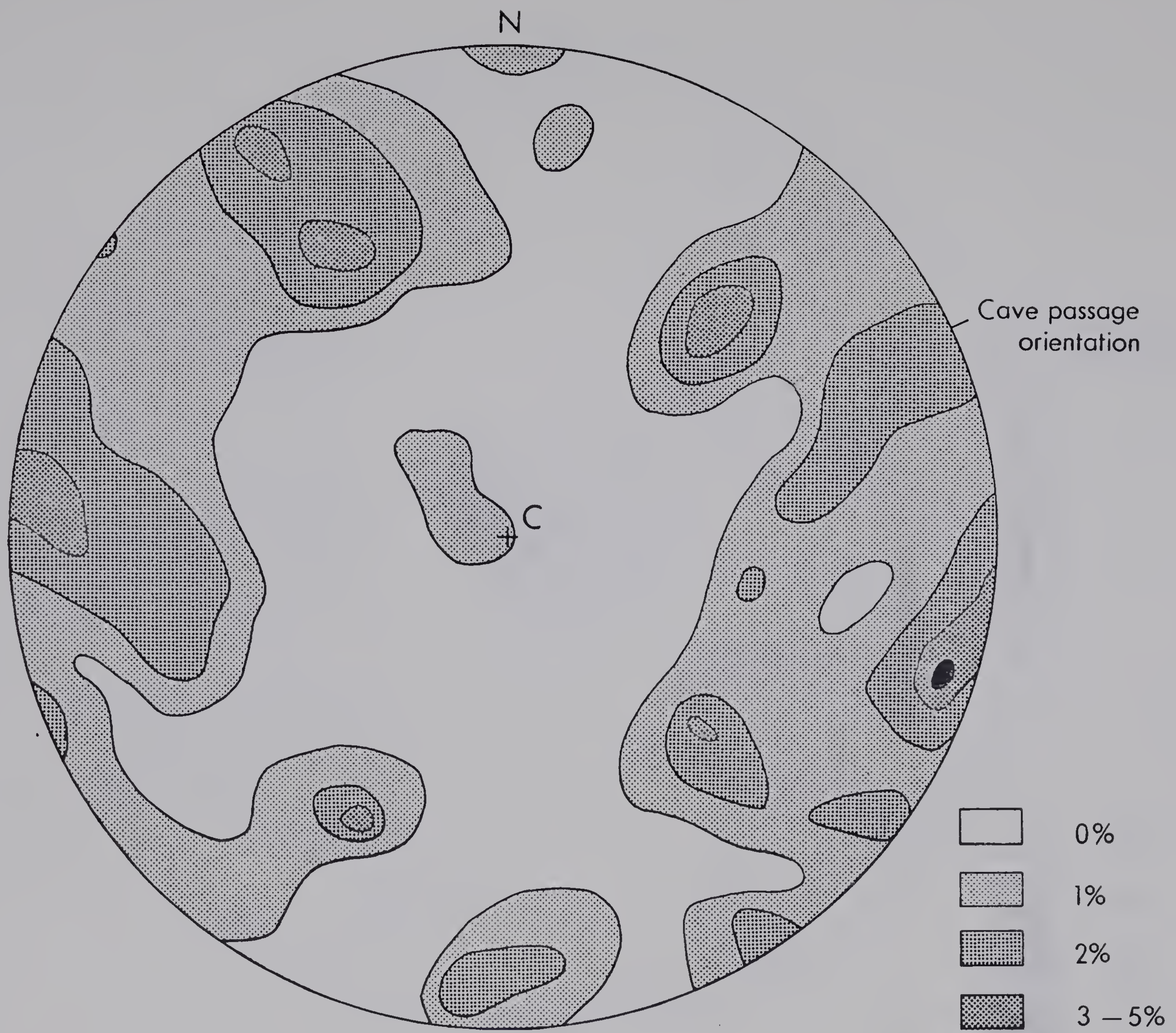
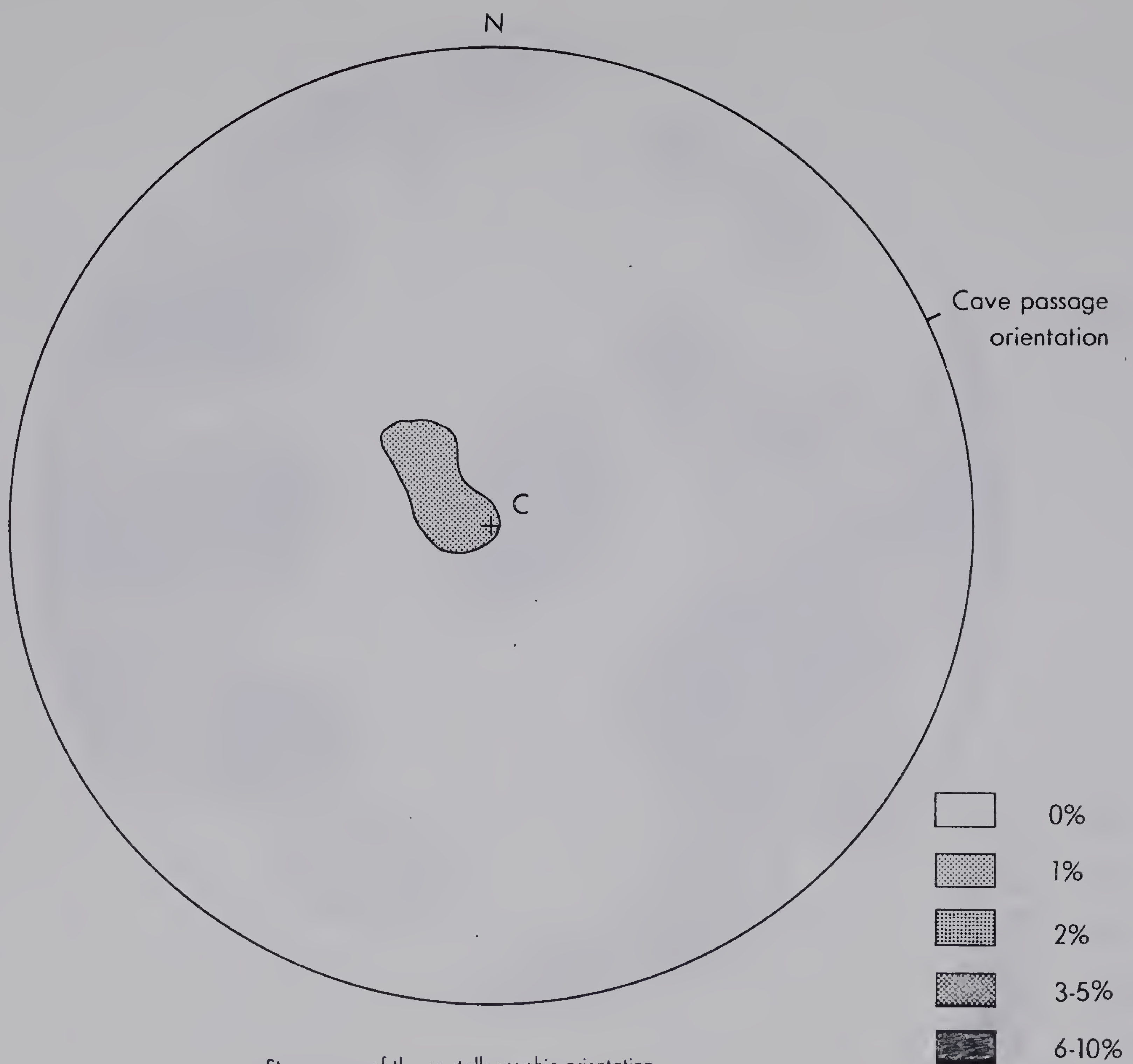
[4] Primitive - The border of a stereogram.



Stereogram of the crystallographic orientation

BLOCK A

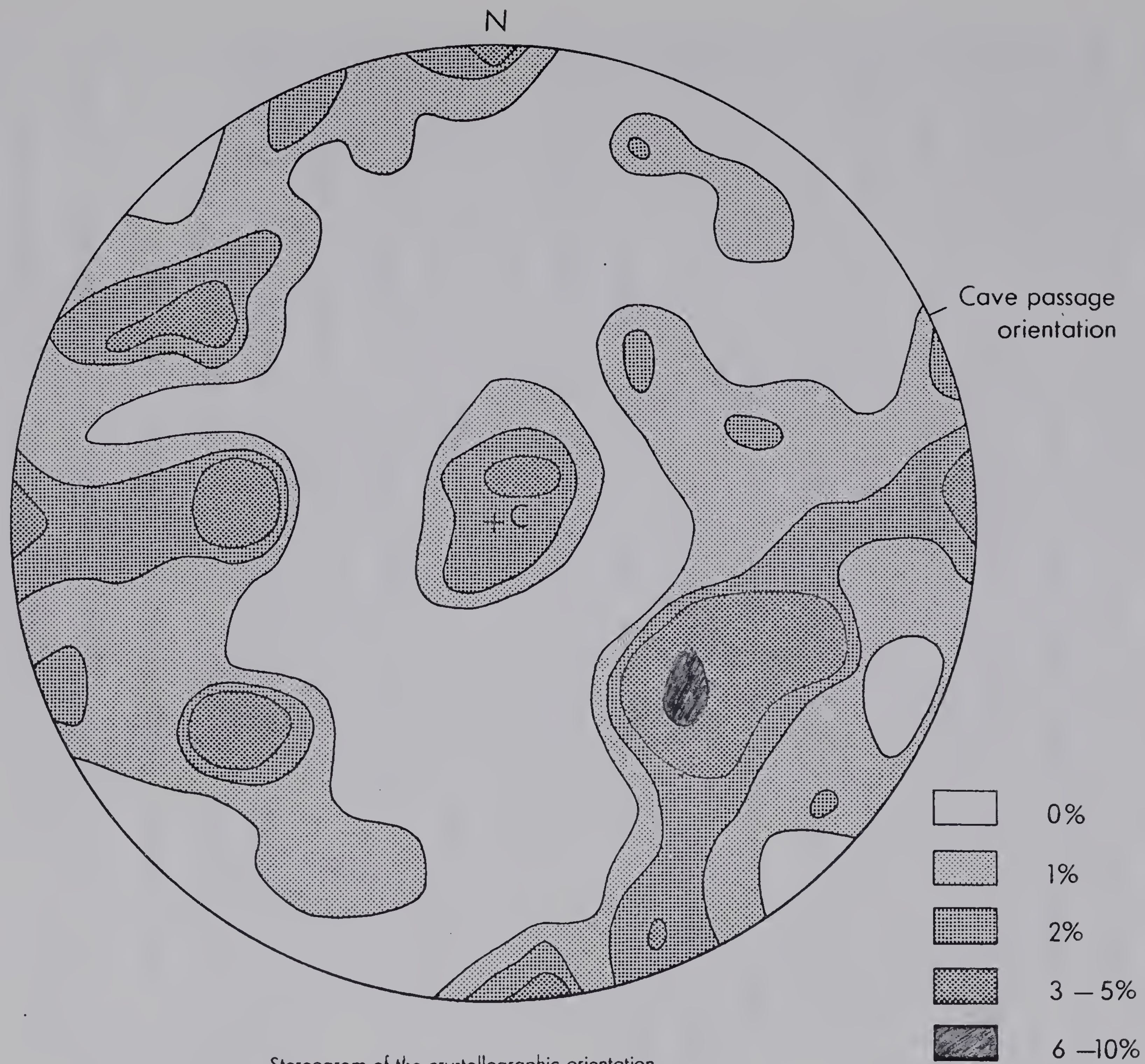
Figure 15



Stereogram of the crystallographic orientation

Upper Block A

Figure 16



Stereogrom of the crystollographic orientation

BLOCK B

Figure 17

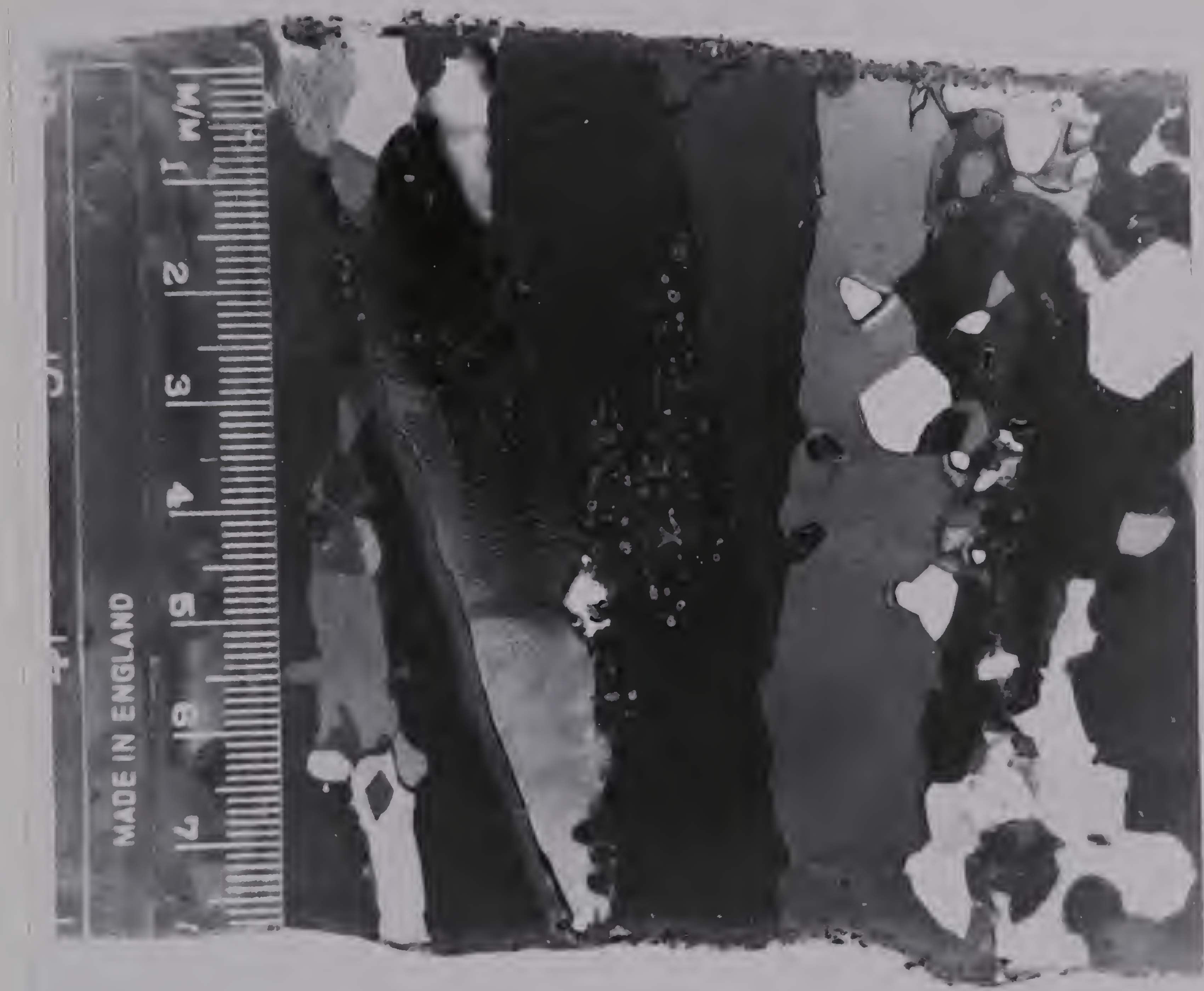
general maxima, the centre of block B is much more heavily represented than that in block A. The former is, however, also composed of the largest crystals which account for 18 per cent of the total number in the block. Moreover, the node is centralized, rather than offset, indicating truly vertical c-axes which have basal plates parallel to the base of the block, i.e., horizontal. Bubble columns are found either along the crystal boundaries or in vertical positions (Figure 18).

The stereogram of block B lacks a strong axis parallel to the cave passage. A cluster is evident in the southwest; a stronger cluster is situated in the western portion of the diagram. While these nodes have no apparent relation to the large crystals, they face that part of the recess abutting the ice-free portion of the passage.

In active glaciers both the largest crystals and the strongest fabrics are found close to valley walls. Here the large shear couple results in crystals parallel with, rather than normal to, the direction of movement (Rigsby, 1960). The fabric diagrams of Coulthard Cave not only indicate dominant crystal orientation normal to the passage bearing but also the long axes of the crystals are vertical whereas in glacier ice the long axes are parallel to the direction of movement. Thus rather than making analogies of ice in the cave to moving ice, the aspect of the freezing of ponded water must be considered.

Ice crystallization

In the initial freezing of ponded water, supercooling sufficient for skeletal growth is often noted (Hallett, 1960). The



North

South

Figure 18. Vertical bubble columns are found either along the crystal boundaries or within crystals in block B.

surface layer of water releases heat by radiation which results in warm convectional air cells. Supercooling soon decreases abruptly in still water owing to the emitted heat of crystallization. Small grains of pure ice form in the surface film to act as seeds for ice formation. The water surface is then covered with a chaotic mixture of crystal sizes and orientations owing to the initial growth of nuclei without influence of one another.

Dendrites, or ice needles, subsequently grow down into the water body (Hallett, 1960) initiating columnar crystals with the basal plates either parallel or perpendicular to the water surface. Basal plates at orientations other than these tow positions shortly become extinct. Crystals whose basal plates are parallel to the water surface may expand outward more quickly than other crystals because of preferential growth on the edges of the plates. When these crystals meet, a growth direction normal to the pond surface is enforced. Otherwise, crystals with vertical basal plates may grow downward faster than crystals at any other orientation. Pounder (1965, p. 74) indicates that crystals with horizontal c-axes will quickly become dominant; Glen (1954, p. 398), Hallett (1960, p. 700) and Shumskii (1964) believe that the c-axes will be vertical: studies of Coulthard Cave ice support the latter contention, as exemplified by the crystal formations in the upper portion of block A, and block B. That the crystals are uniformly orientated in these cases indicates initial growth from vertical or nearly vertical dendrites. If crystallization had initiated from the rock walls, then orientation would depend on the morphology of the rock.

After the initial stage, the conditions under which a specimen is frozen, in particular the freezing rate, determine either the vertical or random crystalline pattern obtained. Shumskii (1964, p. 159) interprets a vertical orientation (centre concentration on the stereograms) as "...slow freezing of a calm water surface...under conditions of intense heat transfer normal to this surface. " In the main, those crystals develop whose direction of greatest rate of growth does not go beyond the limits of the supercooled layer of water. Thus, the thinner the supercooled layer is at the moment of crystal growth, the more uniform the orientation will be. Also, the thinner the layer, the greater will be the passive orienting influence on the basal plates of maximum growth - 'Bertin's law' (Pounder, 1965, p. 100). The only crystals that survive are those whose directions of greatest growth coincide with the direction of orthotropic crystallization[5]. In the process of growth, the remaining crystals are forced out by the more favorably orientated neighboring crystals resulting in a parallel-columnar aggregate. The more the angles of deflection of the c-axes of the crystals deviate from a direction normal to the surface of freezing, the more rapid will be the geometric selection. Crystals with a horizontal axis wedge out, while the few remaining crystals show a corresponding increase in diameter. The ice will then consist of large, very long grains in which the geometric and c-axes will coincide.

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[5] Orthotropic - Mutually restricted growth of ice crystals normal to the surface of the crystallization front.

The length of a crystal will be determined by its width and the amount of water available from below the surface before the dendrite becomes frozen solid, or has its water supply cut off from below (Hallett, 1960; Pounder, 1965).

Depending on the rate and depth of supercooling the deflection of crystals from the vertical may, on the other hand, increase, giving a random orientation. The heat loss through already existing ice grains and the slight supercooling of water associated with it results in spontaneous crystallization. Then not only do previously existing crystals grow, but new crystals appear and become members of the growing aggregate, disturbing its regularity.

The relationship between the roles of ordered and spontaneous crystallization depends on the presence of effective nuclei and on the rate of cooling, i.e., on the magnitude of the temperature gradient. As the temperature gradient increases, the rate of forced, or random, crystallization also increases. This means that nucleation and growth of new crystals will constitute an increasingly larger portion of the ice fabric, the texture of which quickly becomes fine-grained, losing the parallel-fibrous structure and orderly crystallographic orientation it previously had.

More intimately, the shape of any particular ice crystal will reflect the interplay of convection and conduction in the liquid around the basal plates, the distribution of the supercooling influences, and the position of ice inclusions. Conversely, these inclusions may alter crystal growth.

Summary

In order to determine the mode of origin of the ice in Coulthard Cave, samples were removed for laboratory crystallographic research. Determination of the optic axes of the crystals, supplemented by grain size determination indicates the ice formed by the solidification of ponded water. Linear crystals of large dimensions resulted from a thin supercooling layer and strong geometric selection during growth. Protocrystalline aggregates develop from a combination of deep supercooling, the presence of numerous fine bubbles seeding new crystals, and rapid geometric selection. On the other hand, large crystals sampled from a recess in the cave wall grew under two favorable conditions: retarded dissipation of heat and few gas inclusions. Joints in the bedrock were inconspicuous, indicating a lack of high pressure expansion as might be expected if the freezing occurred in passages completely water filled. The significance of these findings will be discussed later.

Chapter IV

Ice Inclusions

Introduction

Natural ice often contains foreign matter which forms inclusions within the ice crystals or between them. In their aggregate state these inclusions may be gaseous, liquid, or solid.

Pure ice as a crystalline substance is very rare. To a great extent the texture of natural ice bodies is due to the presence of other substances in them, while the position of features is, in turn, a function of the formation process of a given ice body. Despite this, only recently has the study of ice inclusions been undertaken (Coachman, 1956; Jonas and Muller, 1969; Matsuo, 1966; Scholander et al, 1956, 1958, 1960; Shumskii, 1964). Theories regarding the formation of greater than atmospheric bubble pressures have been put forward concerning firn snow and glacier ice (Bader, 1950; Coachman, 1956, 1958; Lewis, 1960; Scholander and Nutt, 1960): unpressurized ice of more than a few years in age has received little attention. However, it is this ice, found in some caves, that may provide a natural but simplified means for the study of ice inclusions.

The fate of a foreign body in the past of crystal growth

depends on the following factors:

(1) the magnitude of the forces of surface tension between the body, the liquid, and the crystal,

(2) the resistance of the body to displacement,

(3) the rate of crystallization,

(4) the size of the body.

All the components of ice form a mixture and react to each other only physically, if the chemical erosion of the minor impurities in the liquid phase is ignored. Hence, even during its most disturbing transformations, ice experiences only structural, not mineralogical changes.

Depending on the rate of freezing and the relationship between the forces of surface tension, ice and water, widely distributed impurities can either be captured or expelled. In the case of very slow freezing, such as in the upper portion of block A, or in block B (Figures 10 and 14), all impurities except solid particles the size of grains of sand and larger are expelled by the ice, while every increase in the freezing rate leads to the appearance of more and finer inclusions in the ice (Hallett, 1960).

Gas inclusions

Gaseous inclusions in the ice originate directly from the freezing of water. When water freezes, gases dissolved in it separate completely from the solution. The concentration of air in water increases at the ice-water interface as the freezing proceeds, because ice does not incorporate the air within the crystal lattices (Maeno, 1967, p. 207). When the concentrations of air reach some supersaturation

level, small air bubbles may be formed at the ice-water interface and develop into visible size, at which time they may be captured and become a gas enclosure within the ice. These gas enclosures have been analyzed. The gases dissolved in water which separate during freezing have a new quantitative ratio relative to air. Water equilibriated with air at 0°C contains 2.9 per cent by volume (Frankel, 1964) of dissolved gases having the following composition:

Component	Atmospheric air (Hock et al, 1952)	Dissolved gases (Frankel, 1964)
Nitrogen	78.09	61.54
Oxygen	20.94	34.90
Argon	0.94	1.86
Carbon dioxide	0.03	1.75

The smallest frequent air bubble diameter in block A is approximately 1 mm in diameter with about 34 per cent of the total number of bubbles included in this size group. Of the bubbles larger than 1 mm, some 37 per cent are smaller than 3 mm, 25 per cent are between 3 and 5 mm, the remainder consist of cylindrical enclosures greater than 5 mm in diameter. The nodal size is 1.6 mm. It was noted that the bubbly ice was always quite clear, only rarely showing sediment inclusions.

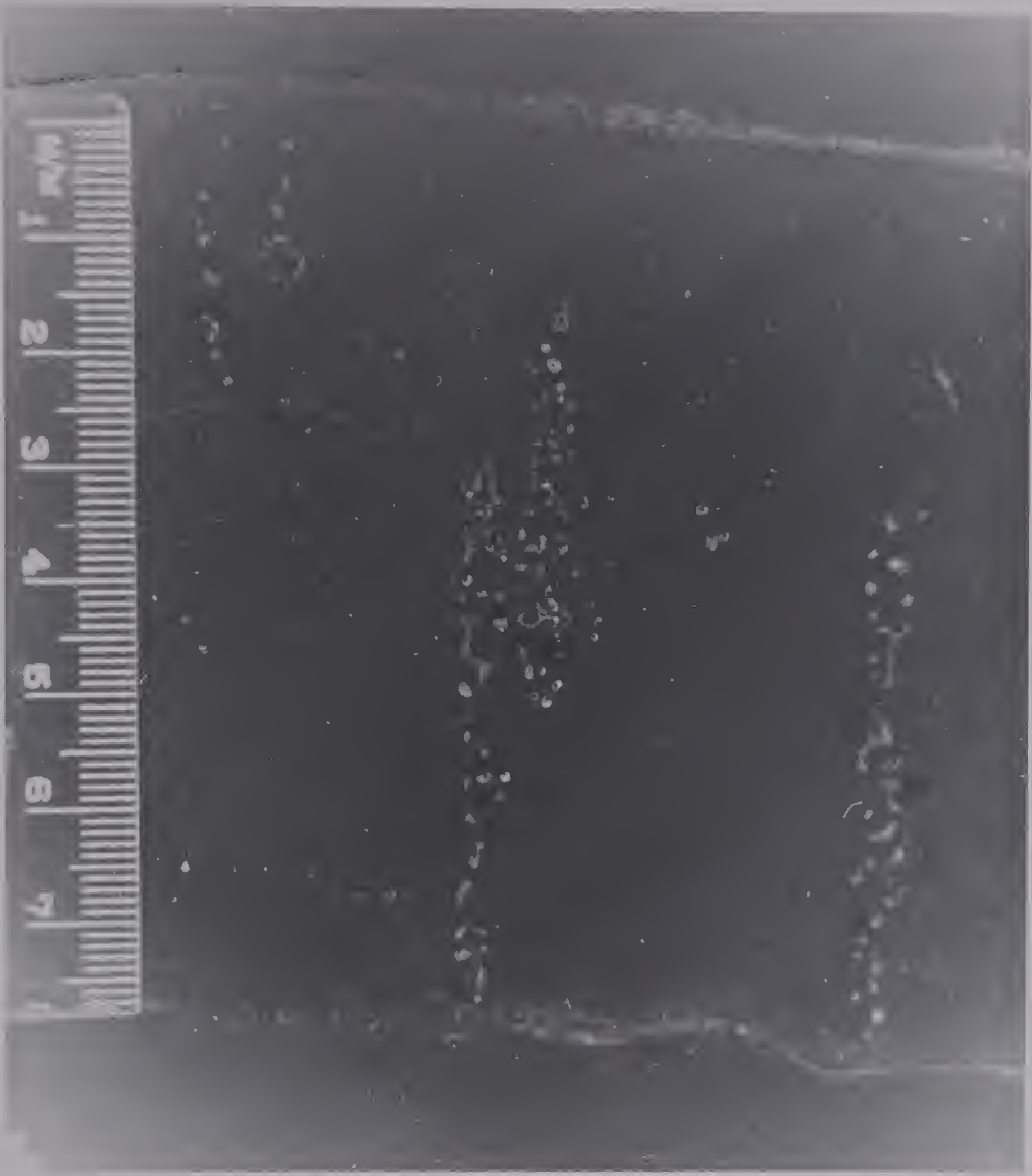
If freezing takes place from below, the bubbles will float quickly upward. At an upper layer of ice, all the air which rises accumulates in increasingly larger bubbles, which are pressed against the ice by hydrostatic pressure (Bently, 1907). If freezing proceeds downward, bubbles simply form and remain at the ice-water interface.

When the rate of freezing is very low, tiny bubbles of air

may remain on the surface of the ice for a long time without growing into it. A slight skin of water is retained between the gas enclosures and the ice; provided sufficient water can replace that which is frozen, the bubble will persist below the ice face (Maeno, 1967, p. 216). During acceleration of the rate of freezing, as discerned in the lower portion of block A, the less than 1 mm bubbles begin to grow into the ice contemporaneously with freezing, often forming nodes of inclusions, which alters the ice from transparency to a milky white colour (Kunsky, 1954, p. 5). Only these small bubbles (maximum diameter 1.5 - 2 mm) remain spherical.

Frequently, instead of a single cylindrical bubble, a chain of tiny spherical or short tubular bubbles will form due to the discontinuous release of air. These chains are elongated in the direction of growth; pear-shaped bubbles are another variant. Most often the broad end of such bubbles abuts the growth face of the ice if the amount of incoming air has decreased during the formation of the particular bubble. If the arrival of air is continuous then the gas enclosures will grow lengthwise, assuming a cylindrical form. Such bubbles indicate the presence and direction of heat flux even more precisely and reliably than do the form and the crystallographic orientation of the ice grains (Bader, 1951, p. 530 ; Shumskii, 1964, p. 175). Thus the bubble chains should be and are vertical in concordant with the petrographic observations (Figure 19).

A characteristic feature of gas inclusions evolving from the freezing of ponded water is that the bubble distribution may be



North South

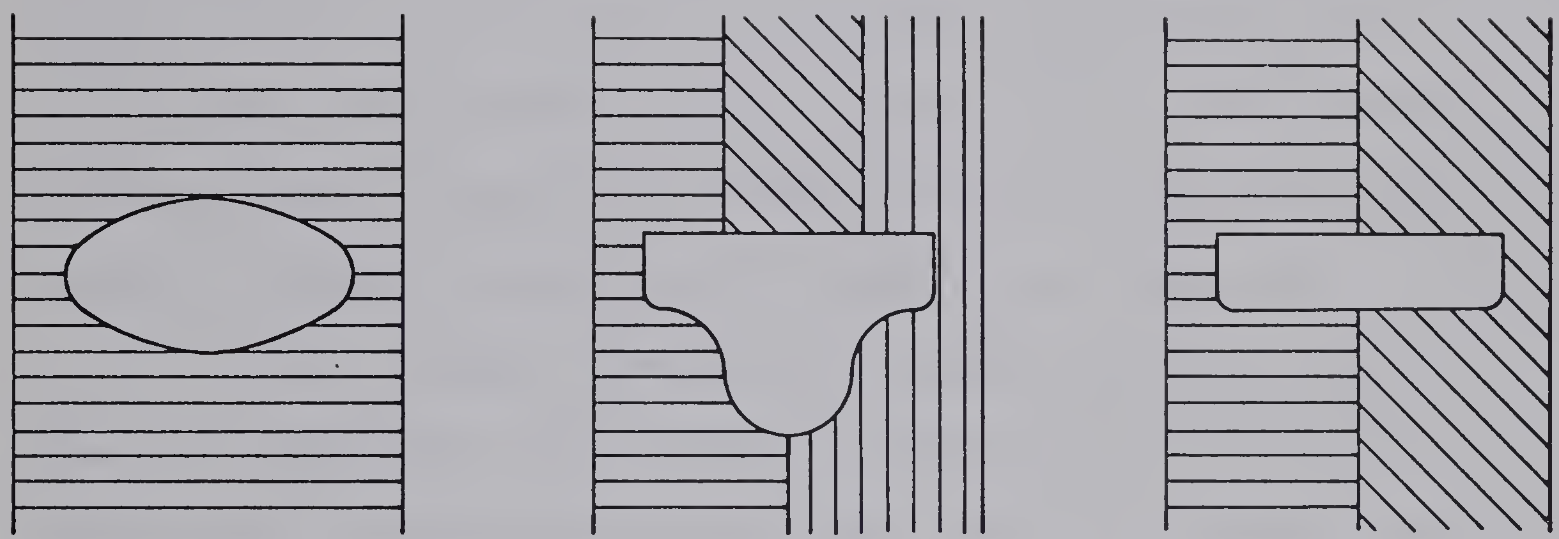
Figure 19. Pear-shaped and small spherical bubbles in vertical chains in block B.

either within ice crystals or between crystals, i.e., independent of the position of the crystal boundaries. If the bubble merely obstructs the growth path of separate parts of the crystal, it will grow into the ice without disturbing the ice texture; however, if it obstructs the entire surface of the crystal, the grain will cease growth and be supplanted by neighboring crystals (Figure 20). The very presence of gas inclusions impedes the growth of crystals, facilitating the development of new crystals which may have a tendency to appear opposite the points of contact of the original crystal and the gas inclusion (Adams and Lewis, 1934).

The air of the bubbles isolated by the growth of ice plays a completely different role than air at the ice-atmosphere interface. The air bubbles play a negative function in the heat transfer process; having an extremely low coefficient of thermal conductivity, the air in bubbles impedes the transfer of heat and reduces the thermal conductivity of the ice.

Bubble migration

When a temperature gradient is applied to ice containing gas inclusions, the inclusions normally migrate slowly toward the warmer temperatures by melting or subliming on the warm wall and refreezing on the cool wall (Shreve, 1967). According to Nakaya (1956) bubbles will migrate under gradients as low as 1×10^{-3} °C/mm. A very tentative temperature gradient of 5×10^{-5} °C/mm has been determined in Coulthard Cave. Yet it would appear that migration at low velocities may still



TYPICAL FORMS OF AIR INCLUSIONS

Figure 20

occur. Stehle (1967, p. 231) calculates that under a gradient of 6.6×10^{-6}°C/mm bubbles in temperate glaciers could have migrated 10 mm since the last glacial advance.

Because a bubble cluster acts as an insulator, the lines of heat flow diverge and the temperature gradient decreases. On approaching the smaller gradient the bubbles would decrease in velocity; migration would, therefore, be from less to more bubbly areas (Stehle, 1967, p. 231). Some areas of even the most occluded ice are greater than 10 mm in breadth of clear ice. It could simply be coincidental that these are vertical, paralleling the bubble trains (Figure 21).

While the age of the ice in Coulthard Cave has not been precisely determined, oxygen isotope findings, as mentioned below, indicate the ice may well be post-glacial, and in fact likely formed only some 4-6,000 years ago. Thus the migration of bubbles probably is not an important function in the forming of present ice texture, or even that of strengthening bubble clusters.

Water enclosures

Tyndall and later observers noted water bags in association with air bubbles in glacier ice (Bader, 1950, p. 443). Even when magnified to 20x no such water enclosures, normally of equal or larger size than the bubbles, and easily visible due to the difference in the index of refraction of water and ice, were noted. As pressure melting is lacking in the 2.7°C

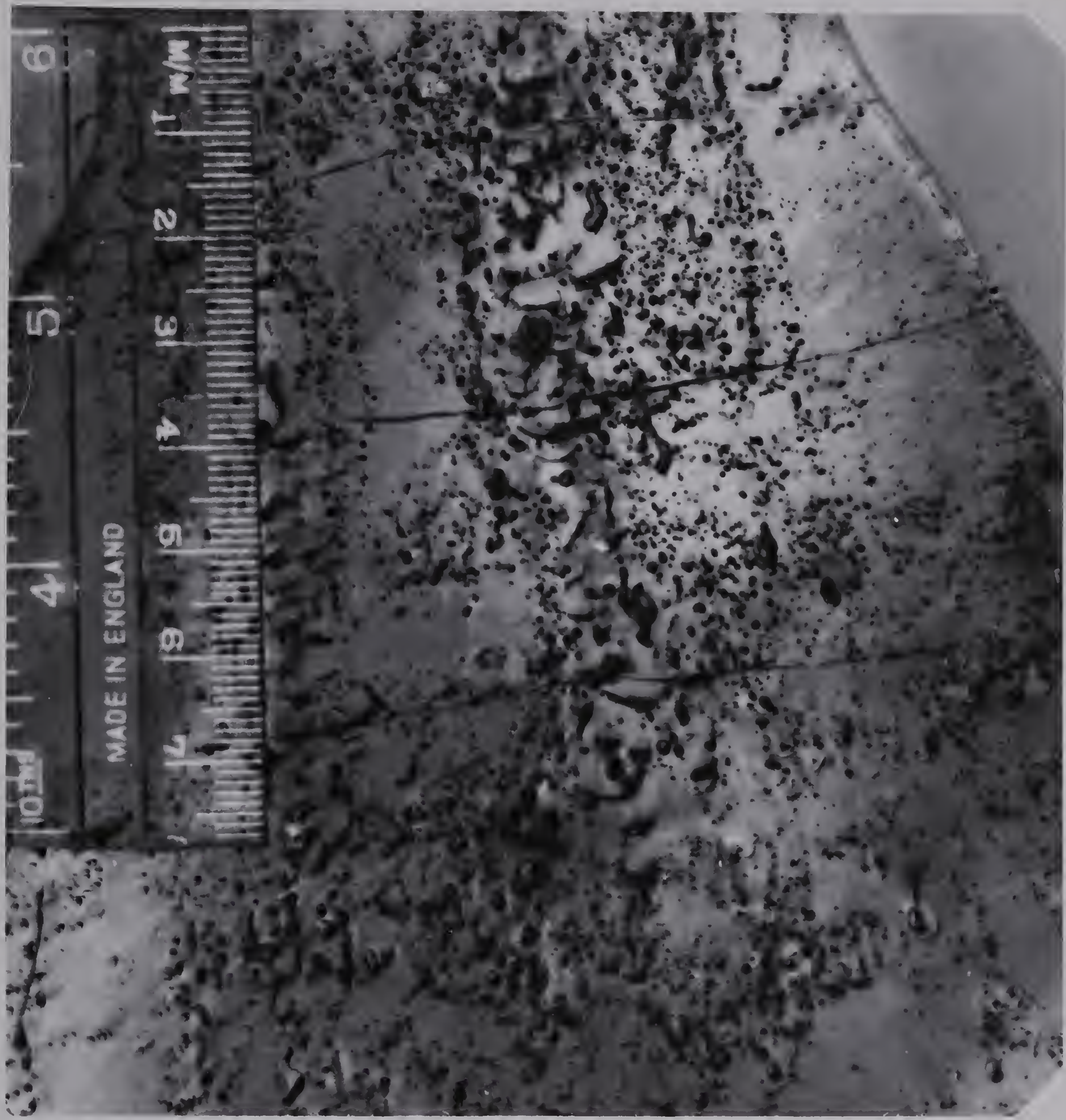


Figure 21. Even in heavily occluded ice bubble-free areas with diameters greater than 10 mm exist.

ice, internal melting could only take place if heat flowed. Bubble regions would, therefore, have to receive heat by conduction. It is known, however, that bubble columns themselves are poor conductors of heat, hence, the presence of water bags is not to be expected in Coulthard ice.

Sediment inclusions

A previous study in Coulthard Cave, on the surface morphology of the ice, indicated that surficial sediment has "...not melted out of the ice"(Goodchild, 1969, p. 126). Yet Stupishin (1959) mentions that water which had been ponded and subsequently frozen contained up to 1.4g/l of sediment, while sublimal ice had 0.2 g/l.

The amount of dry matter in the cave ice was determined by evaporating litre samples. Concentrations noted in the ice were not determined separately, although discrete litre samples were determined from the upper and lower portions of block A, giving 50 and 70 mg/l respectively. Some 20 mg/l of sediment was obtained from block B. The marked reduction in quantity of sediment in block B may be the result of slow crystallization of water in the recess, permitting either precipitation of the sediment, or migration on the ice face. Cort (1963) observed that unless the rate of freezing was high, almost all of the solid particles migrated with the interface, even though air bubbles were trapped and orientated in rows.

Microscopic analysis indicated the sediment to be very fine grained hydrous alluminum silicate - montmorillonite clay - which is

often associated with the weathering of magnesium rich rocks in conditions of less acidity than either kaolinite or illite clay (Moorhouse, 1965, p. 93). Traces of quartz, feldspar and fine muscovite were noted. Concentrations of montmorillonite clay and traces of muscovite appeared in analyses of reduced bedrock material.

Maeno (1967) found that air bubbles were formed between particles and the ice surface, suggesting that the particle surfaces acted as the nucleation site for bubble formation. The small sediment clusters observed in block A did not necessarily occur in conjunction with bubbles: too few were present to make a viable statistical analysis.

Summary

It would appear that the ice inclusions in part depend upon one another and in turn affect the texture of the ice. All are dependent on the rate of freezing. While water under natural conditions contains dissolved gases, and therefore, gives rise to bubbles upon freezing, the slow rate of solidification leads to bubble columns within crystals as well as the formation of large single bubbles. Under a more rapid rate of freezing, sediment adheres to the ice, seeds bubbles, which in turn, form nuclei for new ice crystals, and shortly a polycrystalline aggregate, rather than an ordered texture is formed. Water bags are absent from the cave ice due to a low temperature gradient and the lack of pressure melting.

Chapter V

Isotopic Analysis

Few methods exist for dating ice. Two of the more successful techniques, radiocarbon dating and oxygen isotope analysis, will now be evaluated for use on ice in Coulthard Cave.

Radiocarbon dating

In order to determine the absolute age of gas contained in glacier bubbles so that gas ratios of glacial atmospheres could be determined, Coachman et al, (1958) attempted radiocarbon dating. According to Hemmingsen (1959), permeation of CO_2 through ice occurs at 1 x 10^{-11} cm/sec, at 1 atm and 0°C. He estimated that in 1000 years about 1 per cent of the CO_2 could be dissipated when a 0.5 cm^3 cylindrical bubble under 1 atm pressure containing 0.03 per cent CO_2 is buried 1 m below the ice surface. Thus Coachman et al, felt that the gas enclosed in bubbles should remain relatively unchanged or uncontaminated.

The technique of radiocarbon dating gas enclosures, tested on a Norwegian glacier, required some 5000 kg of ice to obtain sufficient C^{14} for a reasonably accurate count. Even so, only 0.3 g of carbon was available, necessitating two five-day counts. More recent applications of radiocarbon dating ice in situ in Greenland and Antarctica required some 900 kg of ice (Oeschger et al, 1967, p. 939). Unfortunately the

logistics of obtaining a similar mass of ice from Coulthard Cave was well beyond the scope of this study.

Oxygen-18 content

The oxygen-18 concentrations in ice have been used as an indicator of past climatic conditions. Dansgaard (1964; 1969a and b) demonstrated that the O^{18} abundance in the atmospheric moisture depends on the origin of the vapor, the precipitation temperature, and the cooling of the vapor during circulation. It is expected that variations in temperatures between the source of water and the area of precipitation will be reflected by O^{18} values between the sites. This variance has been demonstrated by sampling precipitation (Dansgaard, 1964) and by examining glaciers (Epstein and Sharp, 1969; Macpherson, 1967; West and Krouse, 1969).

As an air mass moves, it preferentially deposits water enriched with the heavy isotope O^{18}, leaving the remaining atmospheric moisture somewhat depleted. Thus precipitation at some later time will be deficient in O^{18} compared to the level at the source. In the case of glaciers, the depletion will be greater with increasing height of the ice mass. When values from large ice caps, eg. Greenland, are similar to those in smaller glaciers, qualitative comparisons in the height of ice and associated climates have been made (Dansgaard, 1969 a and b; Sharp et al, 1960).

In relation to the Standard Mean Ocean Water (S.M.O.W.) datum, values in the order of -35 parts per mille, correlated with the most

recent glacial advance some 17,000 to 13,000 years B.P., have been obtained from deep cores in the Greenland Ice Sheet. These have been attributed to greater amounts of ice over North America which would have reduced the amount of O^{18} being deposited in Greenland.

Krouse has determined the present O^{18} level in precipitation in the Edmonton region and at Banff Hot Springs to be in the order of -20 parts per mille S.M.O.W. This is a significant rise from the Pleistocene value of -40 parts per mille. Ice samples were obtained from Coulthard Cave to determine whether the ice is of recent or glacial age.

Prior to sample collection in the cave, approximately 20 mm of ice was removed from the ice surface to minimize any effects of contamination or preferential loss of light isotopes due to surface melting. Two pairs of samples were then collected from the ice blockage. The ice was chipped from the ice face directly into sterile, moisture free 'Nasco Whirl-Pak' plastic bags normally used for pollen samples, and immediately sealed. A second pair of 5 to 10 cc samples was removed at a depth of 40 mm from the surface.

A single sample was removed from the ice covered floor of the cave (location: Figure 4). Again the sample was obtained at a depth of 20 mm. Finally, the upper portion of an ice stalagmite located near the cave entrance was removed for oxygen determination for present day O^{18}. The six discrete samples were placed in a sealed plastic bag which in turn was put into a container packed with snow in order to maintain constant negative sample temperatures until reaching the laboratory.

The isotopic analysis undertaken by Dr. H. Krouse[1] is based on the standard methods of determination where,

$$H_2O^{18} + CO_2^{16} \overset{K}{\rightleftarrows} H_2O^{16} + CO^{16}O^{18}$$

and K is 1.009 at a temperature of 25°C. The confidence limits are $\pm$ 0.2 parts per mille S.M.O.W.

The data obtained from the $O^{18/16}$ determination are as follows:

Sample	Depth	Parts per mille S.M.O.W.
1	20 mm	- 17.1
2	20 mm	- 16.1
3	40 mm	- 14.9
4	40 mm	- 15.1
floor	20 mm	- 17.8
stal		- 11.8

As mentioned above, the present O^{18} value at Edmonton is approximately - 20 parts per mille, which is more negative than all of the determinations of Coulthard Cave ice. This may be due to the migration of isotopes in cave ice adjacent to the air-ice interface. While a gradient may exist within the ice, the difference in values is not significant until more extensive sampling, especially at depth, can be undertaken to test for any migration.

An isotopic exchange of water percolating through the carbonate horizons above the cave may have concentrated O^{18} resulting in more positive determinations than would otherwise occur. This is particularly

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[1] Dr. H. Krouse, Department of Physics, University of Alberta, Edmonton, Alberta, Canada.

evident in the -11.8 parts per mille value for the ice stalagmite sample, for which water percolated through the limestone roof of the cave. Here also, preferential ablation of the lighter isotopic species may result in concentrations of O^{18} at the surface. Careful sampling of thin layers of ice would be required to test for surface concentrations or migrations.

Chapter VI

Surface Scallops and Sediment

Scallops

Polygonal ablation hollows on snow were known to Heim in 1885; Ratzel, who noticed them in the Alps in 1889; and Hamberg, the author of a classical work on the properties of snow cover in 1907 (Jahn, 1968). Westman, studied the associated dust ridges in 1913, as did Dobrowolski in 1923. Only recently have experiments been carried out on persistent snow patches in alpine regions (Jahn, 1968; Klapa, 1963). Only one major study includes the formation of scallops on ice surfaces (Goodchild, 1969).

The meteorologic condition for snow scallop formation appears to be slow melting, predominantly by sublimation, caused by relatively dry, turbulent air movement near the ground surface from warm to shaded or otherwise cool areas (Jahn, 1968). Generally speaking, the forms are independent of the surface slope and, in fact, it appears to matter little if the scalloped surface is horizontal, vertical or upside down.

Initially melting, or sublimation, begins at points of lower density or surface roughness: a deposit of ice or snow is never perfectly homogeneous (Amstutz, 1958, p. 305). The existence of small depressions such as former gas penclosures may be enough to initiate low velocity cellular convection air movement across the cool surface.

Once incipient hollows are initiated then modification arises from two sources:

(1) changes of the physical variables acting on the medium such as humidity, dew point, insolation, air velocity and direction,

(2) alteration of the internal fabric of the medium.

The latter factor appears not to concern ice scallop development as it does in snow for the surficial ice crystals appear unmodified in laboratory analysis. Variations in crystal size, shown by preferential ablation of the crystal boundaries, within individual scallops and between those at various locations on the ice blockage in the cave, do not result in corresponding morphological changes of the scallops. Goodchild (1969) attributes air movement as the primary determinant of scallop size and shape. While it has been thought that little air movement would occur in descending passages which form 'cold air traps' (Geze, 1965), the movement of air through the main passage of the cave may set up secondary resonance in the side passages. This secondary movement as well as the occurance of heat exchange at the ice/air interface may be sufficient for the formation of convection cells which preferentially sublimate regions of the ice surface to scalloped forms. So as not to duplicate the work of Goodchild (1969), two factors noted on the ice blockages will be discussed. The one is the rate of ice ablation, while the other is the sediment on the ice surface.

A 3mm/yr rate of downward scallop mevement can be roughly estimated from a penny left on the ice in 1968 and recovered in 1970 (Figure 22). It might be conjectured that at this rate of ablation

Figure 22. A penny atop an ice pedistal indicates a rate of ice ablation/ Penny is <u>in situ</u> in the cave.

some 2,000 years would have been required to remove the 5 m of ice still blocking the cave passage above the level of the penny. It is obvious, however, that with only a 2 yr sample period, this estimate may be incorrect by even an order of magnitude. As mentioned above, even within recorded history, several warm and cold periods have been noted, and these would vary the rate of ablation. Experiments carried out in a cold room resulted in a 1 mm daily lowering of an ice surface at a temperature of - 10°C and a constant wind velocity of 4.4 m/s. Samples at the same temperature but subject to wind velocities of less than 1 m/s resulted in negligeable surface lowering over a 7 day period.

Surficial sediment

The surficial sediment is montmorillonite clay. Traces of fine-grained micas, particularly secondary micas, which approach clay in composition, were noted, as well as traces of feldspar and quartz: all are uniformly fine grain clay size and similar to the sediment melted from the ice. The sediment is finely spread over the whole of the scalloped ice surfaces with heavier concentrations occurring near the interstitial ridges (Figure 23). Evidences of splash marks, which would be the result of drops from the roof, or of breaks in the sediment ridges from melt water movement, or ponding, are all lacking. A clean, new plastic sheet 2m in diameter was positioned over the scalloped ice for a period of 26 days in order to make a positive check on the arrival of sediment by drops or, by air movement within

the cave. At the termination of the first sample period, no measurable dust had been collected. A check at the end of an additional 37 days proved inconclusive because of interference by rats (Neotoma cinerea).

Dried white spruce (Picea glauca) needles approximately 12 to 15 mm in length were distributed over three scallops to act as tracers for surface movement or preferred orientation as noted by Jahn (1968) and studies on snow patches carried out by the author. Jahn found that scallops developed on only those arctic and alpine snow patches which persist into spring and summer. These late lying snow patches occur especially in hollows on slopes where wind speeds are reduced, with the result that snow drifts deeply in winter and is slow to melt in summer. Climatic characteristics appear to include slow melting, considerable loss due to sublimation, relatively dry winds often of katabatic or anabatic nature, air turbulence near the ground surface due to the relief of obstacles, a site likely to be shaded, and bare rock surfaces exposed around the snow bank which are able to absorb insolation thereby warming the air creating turbulence (Jahn, 1968, p. 300). Preferred orientation of needles or grass blades may occur within days or weeks on the snow patches studied. In the 63 day interval of the Coulthard Cave study, however, no measurable variation in the pattern was detected.(Figure 23).

Surface movement of particles must occur, however, because marked concentrations of sediment to 20 mm in thickness are found on the crests of ridges and particularly where these crests meet. The largest sediment deposit is some 4 m above the floor of the ice blockage,

partially vertical, partially overhanging. Obviously, therefore, the adhesion of sediment to the ice is strong, yet when sampled, all but that in direct contact with the ice could be easily removed.

Since sediment inclusions within the ice remain stationary ie., are not transported by bubble movement, particles arrive at the ice-atmosphere interface through the downward migration - ablation - of the ice surface. Once at the surface, sediment must proceed by some mechanism to the ridges. Hoekstra (1967) and Itagaki (1967) experimented with bead migration on the ice surface; Ball (1954) put forward a normal trajectory theory of sediment movement (Figure 24).

Bead migration as stated by Itagaki (1967) depends on some unknown force moving or sliding water coated particles across an ice surface. A temperature gradient is mentioned as one possible means of locomotion. Wigley[1] using a temperature gradient considerably greater than that measured in Coulthard Cave between the air-ice and ice-rock interfaces demonstrates that particle movement would be in the order of 6×10^{-6} cm/yr; obviously some orders of magnitude too low for ridge formation since the Pleistocene, and certainly not enough for the period since the end of the Hypsithermal Period. Surface bead migration should also result in accumulations of sediment in the preferentially ablated intercrystal boundaries exposed on the surface.

The simple adhesion of sediment to the ice surface in

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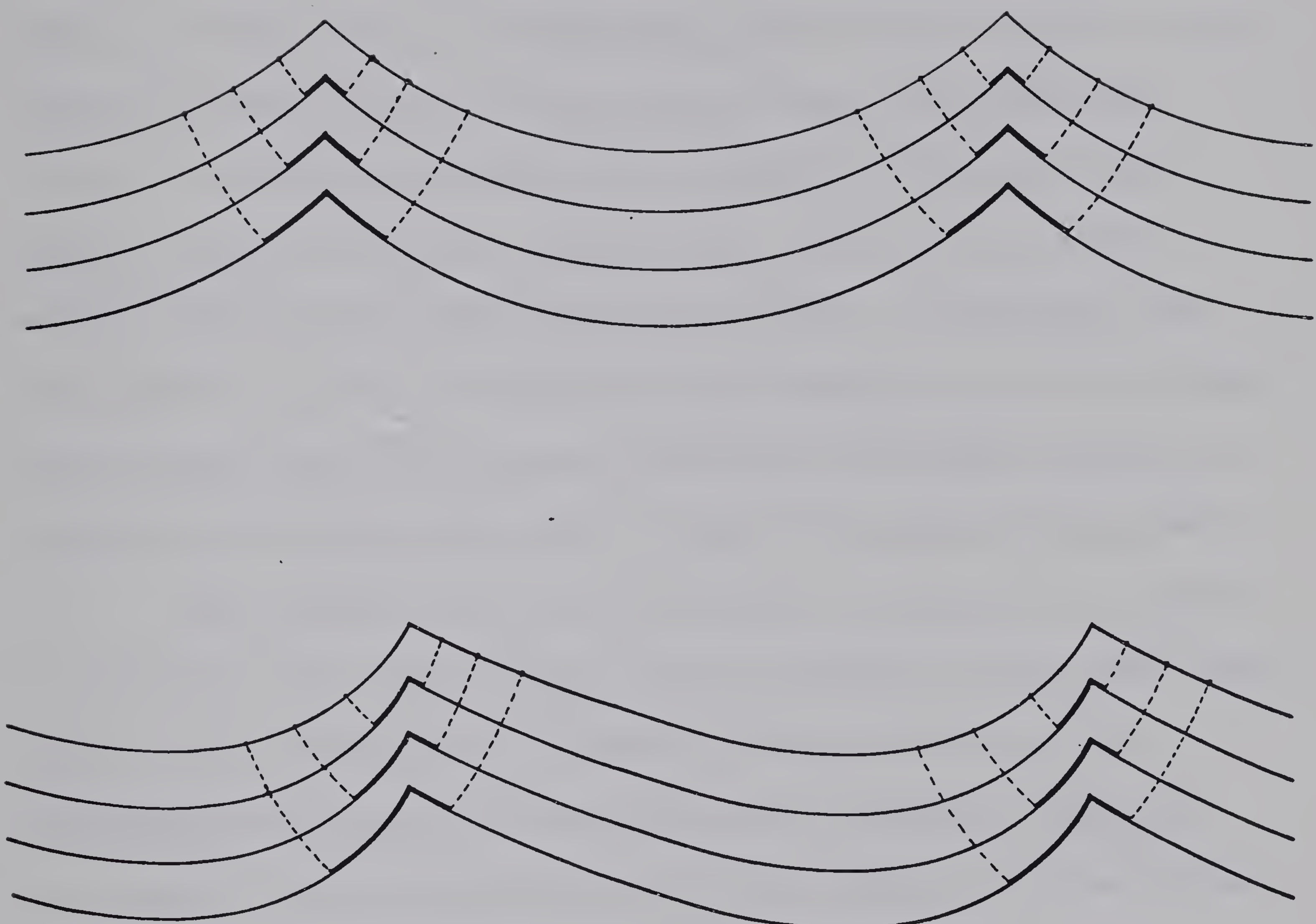
[1] T.M.L. Wigley, Department of Mechanical Engineering, University of Waterloo, Waterloo, Ontario, Canada. Pers. comm.



Scalloped surfaces with sediment

Figure 24

conjunction with a normal trajectory theory overcome the points of contention. The sediment movement is not controlled by gravity, and does not actually move across the scallop surface. Rather, through ablation of the scallop, particles move normal to that portion of the surface on which they rest as the surface lowers. Providing that sediment is uniformly available, those adjacent scallops with the greatest potential ablation distances will also have the heaviest sediment ridges between them (Figures 23 and 24). In Coulthard Cave this would be the vertical scallop on the backwall of the ice blockage, for it could retreat down passage vertically. Retreating scallops are limited to the distance from roof to floor for sediment ridge growth.

The surface of the ice in the cave is scalloped by a series of convectional air cells ablating the ice surface into and maintaining scallop forms. The sediment contained in the ice arrives at the ice-atmosphere interface by scallop ablation. Subsequent 'movement' is by normal trajectory until an interstitial ridge is attained. Those scallops which retreat down-passage have the potentially greatest amount of ice from which to obtain sediment and therefore the heaviest deposits of sediment on ridge crests.

Chapter VII

Conclusions

The initial aim of this study was to determine the mode of origin and age of the ice in Coulthard Cave. During the period of warmer climate known as the Hypsithermal Period, any previous ice formations in the known limits of the cave probably melted. The extent of the passages at that time is unknown, because towards the termination of the warm period colder temperatures influenced the cave climate. Ice blockages formed at depth in the sloping passages by ponding and freezing of seepage water. The ice blockage in the upward sloping passage (Figure 3) developed near the present snow accumulation. Water descending the passage froze, stopping further flow down the passage. These frozen accumulations of water seepage known as 'ice water falls', flows and stalagmites are common to other Rocky Mountain caves (Brown, Ford and Wigley, 1970).

Water containing suspended sediment and dissolved gases ponded in these passages. Because of the latent heat given off by freezing and also because of the high heat capacity of rock, initial freezing of the water was by thin supercooled layers: large vertical ice crystals developed. The largest crystals formed in recesses close to the bedrock. Here gas and sediment inclusions were found within the crystals and were of small size, and therefore had little or no influence on the ice texture

Since evidence of crystal seeding by the rock walls is lacking, development of the crystals proceeded downward from the pond surface.

The establishment of a deep supercooling layer altered the ice texture to that of a protocrystalline aggregate. Nucleation of numerous crystals became possible at the ice-water interface initially because the deeper supercooling had less influence on crystal orientation, but also because increased numbers of gas inclusions seeded crystals. Rapid geometric selection maintained small crystal size. Detectable bubble columns within the aggregate are vertical, indicating a continual temperature gradient normal to the surface of the ponded water.

Recent fluctuations in climate have been damped by the entrance snow-ice deposit in the cave. Though cave temperatures remain negative throughout the year, the ice ablates by sublimation. Thus the ice blockage in the lower portion of the entrance side passage has been slowly eroded. The rate of erosion, at present, is approximately 3 mm annually.

Sediment inclusions arrive at the air-ice surface through a lowering of the ice surface. Adhesion of the sediment to the ice permits even overhanging scallops to maintain sediment ridges formed by a normal trajectory theory of particle movement. For this reason, those scallops retreating down passage amass considerably more sediment than ones moving vertically: the potential distance of the first is unlimited; the second is determined by the passage height. Sediment samples are consistent in texture and composition with bedrock content which shows that sediment within and on the ice was derived from material in suspension when the ponded water solidified.

Determination of oxygen-18/16 ratios indicates the O^{18} content is more positive than present precipitation and, therefore, is not likely to be glacial ice. Rather the ice could have formed from seepage during the warm Hypsithermal Period. The extensive carbonate deposits in the cave and the ablation of ice by sublimation may have altered the original oxygen ratios. Furthermore, migration of isotopic species may have taken place. It is suggested, therefore, that further sampling of all ice deposits within the cave be undertaken. Increment sampling of ice cores could test for isotope migration within the ice. The use of hydrogen deuterium, an isotope of water, or radiocarbon dating could indicate more precisely the age of the ice.

Further research might entail more extensive petrofabric analysis of ice samples to ascertain the depth of former ponds. The findings might correspond to former warm climate conditions on an annual basis. These studies might also reveal more clearly the influence of the heat capacity of the rock walls on crystal growth.

Any or all of these studies would provide more data for a better understanding of the processes which influenced the ice formation.

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